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STRUCTURE AND PETROLOGY OF THE SUNAPEE QUADRANGLE, NEW HAMPSHIRE. By C. A. Chapman

This paper is accompanied by a colored geologic map. Copies of the map (unfolded, in tubes) may be purchased. Price 25 cents. Remittance should accompany order.

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BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
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OUR SHRINKING GLOBE

(Vice-Presidential Address, Section E, AAAS, Philadelphia, December 27, 1951)

BY KENNETH K. LANDES

ABSTRACT

The surface of the earth lies at two levels, the basalt-floored ocean basins and the sial-capped continental platforms. These great segments of the earth's crust are in isostatic equilibrium. Adjustment is made to a shrinking interior by subsidence. First to drop are the heavier ocean basins. This condition of disequilibrium is terminated by subsidence of the continental platforms to approximately the same relative levels as before. The subsidence of the ocean basins may appear, in the uppermost rocks of the crust, as a great downfaulted block (Pacific type) or as a downwarp (Atlantic type). Both the continental and the ocean-basin segments are subject to fragmentation and local subsidence. Examples are basins, geosynclines, grabens, and foundered continental border zones.

The downward movement of crustal segments caused them to crowd together, and the resulting compressional forces were expressed in folds and thrust faults, arcuate mountain systems and island arcs, and geanticlines and swells. Volcanoes were a possible by-product anywhere, but batholithic intrusions were confined to the sial-capped continents.

Possible causes for a shrinking globe include downward solidification of a liquid core, continued cooling of solid rock, and increased density due to pressure conversions. If downward solidification has taken place, the freezing of the crust to its present thickness of 1820 miles would have been accompanied by a reduction of radius of 140 miles and of circumference of 880 miles, using 11.5, the expansion percentage when basalt liquefies, as the contraction factor.

The submarine canyons, some of which extend to depths of 15,000 feet, can be explained as due to subaerial erosion following a subsidence of the ocean basins through a vertical distance of 20,000 to 30,000 feet, which lowered sea level the required distance. Subsidence of continental areas, such as the Gulf Coast of North America, to even greater depths is known to have taken place. The subsidence of the continental platforms restored the isostatic balance and brought an end to the canyon cutting and to other topographic changes, such as the truncation of volcanic cones and the cutting of terraces, that were taking place on the exposed sea floor.

Differential subsidence also explains the glacial climates. A lowering of sea level of 3000 feet would be adequate; this could be brought about by a subsidence of the deeper parts of the ocean basins of 5000 feet. During the Pleistocene there were three subsidences of at least 5000 feet and one of perhaps 25,000 feet.

The lagging behind, in time of subsidence, of strategically placed crustal blocks would create temporary land bridges. Widespread inundation of continental areas took place due to (1) rotation of the platforms during subsidence; (2) drag of the marginal belts due to ocean-basin subsidence; and (3) downwarp (basins, geosynclines) due to fragmentation and differential subsidence of the continental segments.

The normal, isostatically balanced, condition is as today, with the continental platforms largely emergent. The sedimentary sections with their contained fossils available to us by being above present sea level can only be considered brief, irregularly spaced, and totally inadequate samples of the sedimentation and evolutionary history of the earth.

INTRODUCTION

The contraction hypothesis is one of the more venerable of the geologic philosophies. Its history has been reviewed by Willis (1929), Daly (1933, p. 250), and others, so I will not attempt to trace it here. Among its earlier proponents were Dana, LeConte, and especially Suess. Although contraction is not generally accepted today, the hypothesis still has its advocates. Hobbs (1944) uses it to produce island arcs. Jeffreys (1931, p. 437) states that in spite of the lack of general acceptance of the

contraction theory "there certainly are no grounds for accepting any other." De Lury (1945, p. 2) "accepts thermo-contraction as providing the most promising causes of earth deformation."

One possible reason for the disinterest of the majority today in contraction is that there has been too much skin-of-the-apple or coated balloon (Bucher, 1924, p. 276) analogy. There are differences between the crust of the earth and the skin of an apple, and these differences render the analogy a questionable one. The earth's crust will break into fragments and subside instead of crumpling. It is not homogeneous, either laterally or vertically, for it consists of discontinuous sial of variable thickness superimposed upon continuous sima. Earth contraction is therefore differential and localized because of differences in skin weight, thickness, and conductivity.

Nölke (1933, p. 172; 1939), a contemporary contractionist has examined critically the objections to the contraction hypothesis and concludes that "they have no conclusive power, as they are based on suppositions which either are not confirmed by the observations or are of another nature."

I favor contraction because it can be adapted to explain so many puzzling features in geology that it becomes a veritable panacea. Given a choice of several hypotheses, none of which can be either proven or disproven, why not choose the one that explains the most?

Outward Effects of Contraction

Crustal Mechanics

I believe that the reduction of volume in the earth's interior has resulted in sporadic but continued sinking of crustal segments. The emphasis is on vertical movements, and *all these vertical movements are downward*. Horizontal movements have been secondary, owing to crowding of the segments (Bucher, 1950, p. 498) as they moved downward.

The largest segments are the continental platforms and the ocean basins. The only difference between the two is that the former are capped by sial, a relatively thin veneer of granitic rock. Beneath the sial on the continents, and directly beneath the ocean basins, is the sima, the uppermost part of the basalt-dunite crust. We have good reason to believe that the Atlantic as well as the Pacific and Indian oceans is underlain to a large extent by sima (Wilson, 1940, p. 301; Wilson and Baykal, 1948, p. 51; and Ewing and Press, 1950). Because of the greater density of the ocean-basin segments, they tend to lie at a lower elevation than the sial-capped continents. The former can be considered megagrabens and the latter megahorsts, defining a horst as a block that has been faulted downward a lesser distance than the surrounding grabens.

Kuenen (1950, p. 129–132) has pointed out that the Atlantic, Pacific, and Indian ocean floors lie at approximately the same elevation, except for a few excessive deeps in the Pacific that give that ocean a greater average depth. The difference between the average levels of the ocean basins and the continental platforms is 2½ miles (Shepard, 1948, p. 314).

As a general but perhaps not invariable rule the adjustment to fit a contracting interior was made by the heavier oceanic segments sinking first, followed shortly (geologically speaking) by the dropping of the continental blocks to the same general *relative* level as before. Isostasy has been operative on the continental platform-ocean basin scale, but I doubt if it has been important in segments of much lesser magnitude.

Besides the sinking of segments of continental and ocean-basin size there has been widespread fragmentation of these segments, with different fragments dropping widely varying distances. Examples will be given in a later section. The surface expression of a downsinking, whether it involved a relatively small fragment of a segment or an ocean basin, could be either a graben or a downwarp, depending upon whether the rocks at that level broke or flexed.

Some rotation of the sinking crustal fragments took place. The random size and orientation of the dropped blocks are paralleled on a much lesser scale by the Mackinac Straits megabreccia which I have described elsewhere (Landes, 1945, p. 123–153) as due to the collapse of the roof above large salt caves.

The downwarp of the Atlantic basin dragged the surface of the basement rocks of the continental platforms down to great depths in places. According to Ewing *et al.* (1946, p. 932)

seismic measurements of the depths to basement in shallow waters along the western margins of the Atlantic showed depths off the northwestern coast of South America as great as 22,700 feet. The fault-bounded Pacific Basin does not show similar downward drag of the edges of the continental platform; however, the foundering of the continental margins about the Pacific (to be discussed presently) may have been encouraged by the subsidence of the Pacific Basin. Likewise a possible explanation for the deep trenches off the island arcs of the western and northern Pacific is that the sea floor may have been dragged down with the last subsidence of the neighboring sial-capped crust.

Because the planes bounding the subsiding masses converge toward the center of the earth, lateral compression is an inevitable by-product of the sinking of the segments. A common concurring feature is volcanism, and in the sialic segments intrusive igneous activity often took place.

The intensity of the diastrophism in the continental areas was a function of the distance the oceanic blocks dropped; the blocks were deformed by the subsidence of the sial-capped segments. Normally the megagrabens and megahorsts made alternating drops of modest magnitude, but every now and then large areas in the ocean basins subsided to unusual depths, say from 5000 to 10,000 feet below previous levels, and once, to produce the deepest of the submarine canyons, to depths of 20,000 to 30,000 feet. Before you shy away from these figures with complete horror, let me remind you that the vertical displacements of many known continental features are of that order of magnitude!

When these major changes occurred, a diastrophic revolution took place. The belts of weakness, such as the geosynclines, were crumpled and thrust-faulted. Batholiths were injected into the upper zones of the sialic crust, and volcanoes broke out at the surface. At the same time, the lowered sea level produced glacial climates, and mountains were carved from the uplifted areas by differential erosion. The great mountain-building revolutions of the late Proterozoic, Paleozoic, and Tertiary fit into this scheme of things. The Tertiary diastrophism has continued into the present, owing to a series of subsidences during the Pleistocene. In North Africa and the Middle East are folded artifact-bearing beds; in California the late Cenozoic diastrophic revolution is still in progress (Gilluly, 1949).

Evidence of Vertical Movements

As Suess (1904, p. 603–604) has stated, subsidence has "left its traces everywhere." Moore (1936, p. 1807–1808) has pointed out the importance of subsidence in continental tectonics. Examples include the structural basins, such as the Michigan, Illinois, and Permian (West Texas) basins, the geosynclines, the marginal embayments, and the grabens. The Michigan Basin has subsided an estimated 14,000 feet (Cohee, 1948), the Appalachian geosyncline 30,000 feet, and the Gulf Coast embayment (perhaps better described as the north flank of the Gulf of Mexico Basin) somewhere between 30,000 feet and 70,000 feet. The most notable grabens are those that make the Rift Valleys of Africa.

The same evidences of subsidence can be found in the ocean basins, only here the downwarps do not become filled with sediment. The ocean basins themselves appear to be great subsidence features. In some areas, notably the Atlantic, the sinking was carried out by downwarping. Elsewhere, such as around the borders of the Pacific, most of the subsidence appears to have taken place by faulting. Shepard (1948, p. 193–194) believes that most continental slopes "represent a zone of recurrent faulting at the juncture between the light continental and the heavy oceanic segments of the earth's crust." The continental slope off the California coast is a great escarpment rising as much as 10,000 feet above the ocean floor. This escarpment, were it exposed, "would probably be as impressive in declivity as the great eastern slope of the Sierra Nevada" (Shepard and Emery, 1941, p. 21).

Superimposed upon the ocean floors are downwarped basins and deep troughs which may be either elongate sags similar to geosynclines or downfaulted graben blocks. There does not appear to be any more reason to ascribe downward flexing of basins and geosynclines to the weight of the sediment than to credit graben faulting with the same motivation. As Jones (1950, p. 110) has stated, following an

opinion expressed by Jeffreys, "geosynclines developed because of a 'sinking feeling' in the crust. Once sinking occurred near a continental area, filling in with sediments *pari passu* would follow." Sea-floor basins along the California coast are actively subsiding and are being loaded with sediment today, but farther offshore are other likewise actively subsiding basins *not* being loaded with sediment.

Other continental slopes and coast lines besides the Pacific appear to be fault scarps. According to Wiseman and Sewell (1937), the whole of the northern part of the Arabian Sea, bordering Africa, Persia, and India, is surrounded by faults. An undersea photograph of an escarpment that can be interpreted in no other way than due to faulting was recently published (Jordan, 1951, p. 1989). The camera was poised a few feet above the edge of the West Florida escarpment, which drops the ocean floor from 3000 to 6000 feet. The gradient on the scarp face is as much as 35° and may be even more.

Submerged topography highly suggestive of basin and range has been described by Shepard (1948, p. 176–178) and Shepard and Emery (1941, p. 9). Examples cited occur on the subsea Blake Plateau (south of Cape Hatteras), on the continental slope west of the Mississippi delta, and along a zone 160 miles wide off southern California. The epicenters of the marine earthquakes off California tend to concentrate along the submarine scarps showing that movement is continuing (Clements and Emery, 1947, p. 312).

Many sea-floor trenches have all the features of grabens or rifts and have been so designated by investigators. Examples are the Gulf of California, the Gulf of Aden, Red Sea, Bartlett Trough and other trenches in the Caribbean, and the Aleutian Trench. The last-named is 2200 miles along by 50 to 100 miles wide; its maximum depth exceeds 25,000 feet (Murray, 1945). Still deeper are trenches off the southern Japanese Islands and the Marianas, and off Mindanao in the Philippines. These extend down to depths in the neighborhood of 35,000 feet.

The fragmentation and subsidence of marginal belts of the continental platforms have been advocated by many geologists (Barrell, 1927). These inferences are based in part upon the need, as a source of sediment, of a land mass in an area now covered by water, and in part upon petrological and geophysical evidence to the effect that sialic-capped crustal segments now lie beneath oceanic waters, in places at great depth. The best example of the latter is that part of the southwest Pacific lying to the west of the so-called "Andesite Line". This is a line drawn between the volcanoes that yield basaltic lavas and those that produce andesite, a less basic, more continental type. Hess (1948, p. 422) states that the Andesite Line is a major structural and physiographic boundary as well as a petrologic boundary. It separates the North Pacific Basin province from the western island-arc provinces. Deep troughs lie on the ocean side of the Andesite Line (Chubb, 1934, p. 292). Several have referred to the ocean floor west of the Andesite Line as the foundered continent of Australasia (Stearns, 1945, p. 615).

Other areas suggested as foundered continental segments include the Mediterranean (Suess, 1904, p. 603–604), Cascadia (Eardley 1951, p. 62–67), and Arctica (Eardley, 1948, p. 432). Kuenen (1950, p. 168–169) notes that the present continental terraces are young—Mesozoic and Cenozoic. He wonders if the Paleozoic terraces have all foundered, so that they now lie beneath the deep sea waters.

The presence of thousands of feet of reef rock, built entirely by organisms of the shallow-water type, on Pacific islands, such as Funafuti and Bikini, and on Atlantic islands such as Andros in the Bahamas, is further evidence of a subsiding sea floor.

It has already been stated that vertical movements are still taking place in California. In many other places such movements have been quite recent. Willis (1930, p. 270) refers to discoveries in the Great Rift Valley of Africa, by an archeologist working with a geologist, which led to the conclusion that early man may have witnessed the later developments of the rifting. According to Moody (1951):

> "The areal distribution of the Citronelle formation yields suggestive evidence that an extensive and perhaps profound subsidence of the northern portion of the Gulf of Mexico occurred in latest Pliocene or earliest Pleistocene time."

The so-called "deep-focus" earthquakes (Umbgrove, 1948, p. 150) result from slippage along fault surfaces at depths ranging from about 200 to 400 miles below the lands bordering the Pacific basin. Daly (1940, p. 400–401) points out that some interpret this to mean that the low strength necessary for isostatic adjustment cannot begin until below the maximum depth reached by these earth movements, which are indicative of a brittle crust. According to Hess (1948, p. 436–437) both the deep-focus quakes and those originating at more moderate depth in the western Pacific appear to occur along a plane inclined about 45° westward from the tectogenes. This means that the foci of such quakes may lie beneath the mainland. It would appear that the Pacific Basin segment is subsiding today. I believe that other evidences of contemporary diastrophism visible in the circum-Pacific area can be ascribed to the pressure exerted against the bordering continental segments by the subsiding Pacific Basin.

The contraction concept explains the virtual absence of deep-sea sediment upon land areas, but does not militate against land areas becoming the locale for deep-sea deposition. Surfaces can sink to great depths, but cannot rise to great heights. A possible exception is in the Barbados Islands where typical oceanic oozes have been found. According to Schuchert (1932, p. 881) the floor on which the ooze was deposited must have sunk "6000 or even 10,000 feet and later rose to 1100 feet above sea level." The sinking is readily explained; the apparent rise may have been due to subsidence of all the surrounding area, or it may be an actual rise due to compressive forces.

Evidence of Horizontal Movements

The crowding of the blocks due to convergence toward the center of the earth produces horizontal pressures that are relieved by upwarping, folding, and thrust faulting. This crustal shortening takes place in ocean and continental segments alike (Lees, 1951, p. 108–109). On the ocean floors, ridges, swells, and island arcs result. On land the squeezing produces upwarps, folded belts, and overthrust sheets. Some compressional sea-floor features connect directly with similar structures on land. Thus the Lower Antilles lie upon a geanticline that connects the Greater Antilles with the east-west ranges of Venezuela, according to Hess (1938, p. 94).

The evidence that the orogenic phase of mountain building has been the intense deformation, accomplished by both folding and thrust faulting, of geosynclinal belts is so well known that it does not need repeating here. It is also generally agreed that geosynclines filled with sediment and warped downward deep into the upper part of the crust are potential belts of weakness. If, then, horizontal compressive forces are exerted in the crust, these forces could be satisfied by the pleating of the crust along such vulnerable belts. The net result would be a shortening of the earth's circumference. This is demanded by any contraction theory, and the visible shortening that has taken place gave rise to the contraction hypothesis in the first place.

One reason for the current disrepute of contraction has been the tendency of some students of mountain structures to estimate a crustal shortening far in excess of the maximum shortening considered possible by any contractionist. Recent estimates of the amount of shortening in the Alps alone range from 100 to 1800 miles! These estimates provoke the following comments:

(1) The wide range in the estimates above make their accuracy questionable. It is possible to iron out folds so as to determine the shortening within close limits, but estimating the displacement of overthrust faults borders upon guesswork.

(2) If the shortening amounted to say a modest 200 miles, what became of the same 200 by 90 by 25 cubic miles of rock that this shortening must have thrown up in the Alpine area? Certainly the volume of the mountain mass today, plus the volume of the post-orogenic sediments, is far below the total required.

(3) If this crustal shortening did not take place over a contracting interior, where is the area that had 100 to 1800 miles of blanket pulled off it?

All this does not, however, solve the problem of over-contraction. I like Jeffreys' (1931) sug-

gested solution. He believes that "crustal shortening and the horizontal movement measurable at the surface are two very different things". Rocks uplifted by a modest crustal shortening may spread laterally under their own weight to produce nappes and klippes which give a false idea of the magnitude of the horizontal reduction. Substantiation of this idea on a small scale has been found by Harrison and Falcon (1936) in southwestern Iran. There when limestone mountains were piled up to heights of 2000 feet or more they have tended to collapse outward by gravity, producing structures suggestive of the Alpine flysch and nappe. De Lury (1945, p. 9) explains the apparent excessive shortening as due to "autotraction", contraction caused by drag in a thrust sheet into which viscous magma has been forcibly injected.

The island arcs and arcuate mountains are ascribed by Hobbs (1923; 1944) to underthrusting of the ocean floor against the continental margins; the cause of the push is the settlement of the ocean floors (due to thermal contraction) with "the resulting reduction of their superficial areas." Island arcs are notable are their foredeeps and their active volcanoes (Umbgrove, 1948, p. 145). The volcanoes are built upon from one to three parallel geanticlinal swells. The western belt of arcs in the Pacific, extending from southern Japan through the Ryukyus and Formosa to the western Philippines, is thought by Hess (1948, p. 418) to be mid-Mesozoic. The diastrophism producing the eastern and southern belt started in the late Cretaceous and Tertiary and has continued spasmodically to the present.

Hess (1948, p. 442) has termed the deep trenches on the ocean (convex) side of the island arcs "tectogenes". These show strong negative gravity anomalies, and it has been suggested that they have been formed by downbuckling of the top of the crust (Vening Meinesz, 1934). The ocean floor on the ocean side shows a positive anomaly, and the geanticline on the concave side produces an even stronger positive anomaly.

Lateral compression with resultant crustal shortening in the ocean-basin segments is not so obvious as on the continental platforms. For one reason the rock is more homogeneous; wedges of more vulnerable rock do not extend down into the crust beneath the ocean floors. Perhaps most of the shortening of the earth's circumference has taken place within the continental platforms. In addition to the island arcs are other positive topographic features of the ocean floor that are most easily explained as upwarps. Examples are the Mid-Atlantic Ridge and the Hawaiian Swell. The higher points of both are volcanic peaks, but they rise from much larger features. The Atlantic Ridge can be traced from Iceland to the Antarctic; it rises an average of about 6000 feet above the adjacent ocean floor (Shepard, 1948, p. 280-281). The Hawaiian Swell is a gentle rise 600 miles wide and 1900 miles long (Betz and Hess, 1942, p. 101).

Status of Ocean Water in Contraction

Regardless of whether the ocean water originated from without, owing to the condensation of an envelope of water vapor which surrounded the cooling earth (Nutting, 1943), or came from within, as a normal product of magmatic activity (Rubey, 1951), and in spite of its importance geologically and climatically, the presence of ocean water is entirely fortuitous as far as my concept of megatectonics is concerned. It has had no effect, except to add some weight to the topographically lower segments. During times of maximum relief of the segment surfaces the oceans were much withdrawn, concentrated in the areas of lowest elevation. During times of minimum relief the oceans spread far and wide over the earth's surface. In other words the oceans are about as significant to the contraction hypothesis as a drop of rainwater on a teeter-totter, and behave similarly.

Possible Causes of Contraction

Suggested Causes

My concept of contraction calls for a shrinking globe, with the crust fragmented and subsiding differentially. I am not greatly concerned whether that shrinkage is caused by (1) downward solidification of a liquid core; (2) continued cooling of solid rock; (3) increased density owing to pressure conversions; or (4) increased density for any other reason. Jeffreys (1950) is an

advocate of thermal contraction after the crust has solidified upward from the top of the liquid core. Bridgman (1951) has found some evidence that high pressures may produce polymorphic transitions with a net change in density; the implication could be that pressure might convert material deep in the earth to another crystal form of lesser volume and greater density.

Downward Solidification

Personally I favor contraction due to downward solidification of the liquid interior, but I do not insist upon it. I prefer this hypothesis simply because it appears to explain better the sequence of events during that part of earth history of which we have some knowledge. Fundamental to this idea is a once-liquid globe that has solidified up to the present to a depth of 1820 miles. Presumably with time the entire sphere will become solid, as possibly has occurred with the smaller planets. The solidification of a crust 1820 miles thick would be accompanied by a reduction in radius of approximately 140 miles and in circumference of 880 miles. These figures are obtained by using 11.5, the expansion percentage when basalt liquefies (Yoder, 1951), as the contraction factor due to solidification. Even if too large by as much as 50 per cent, which is unlikely, the magnitude of the reductions is adequate for the subsidences and crustal shortenings called for in this paper. No attempt is made to include the additional reductions which must have followed due to post-solidification cooling.

So far as I am aware there is no factual evidence that contravenes this hypothesis of cause of contraction. At least four points have been raised in opposition to this concept, but I do not believe them to be irrefutable.

(1) The boundary at depth 1820 miles is a chemical one, between an inner core dominated by iron and nickel and an outer crust of silicates. However, this major discontinuity could equally well be solely physical, the interface between liquid core and solid crust, with the chemical nature of the materials immediately above and below the discontinuity the same (Turner and Verhoogen, 1951, p. 349).

(2) The crust solidified upward from the core (Jeffreys, 1950, p. 117–118). This idea is based upon (a) the existence of a chemical boundary at the core, and (b) an assumed rate of temperature increase downward at the time of solidification; the actual rate might have been much higher.

(3) Once the crust had solidified it would act as an extremely efficient insulating blanket, preventing the escape of any further heat. I am indebted to Professor J. M. Cork of the Department of Physics at the University of Michigan for calculations that led to the conclusion that it is not unreasonable to assume sufficient heat loss from a liquid core by the conduction process alone to account for continued solidification downward. Furthermore, the entire heat loss does not have to be solely by conduction through the solid crust, for there are leaks through which heat escapes. Every volcano is a heat tap, with heat escaping by ejectment of both gas and lava. The fissure flows of the geologic past, such as those in the Tertiary, provided means for escape of great quantities of heat. The intrusions, especially those on batholithic scale, transfer heat surfaceward. Every hot spring, regardless of source, drains heat from the earth's interior.

(4) The heat escaping from the earth's surface today is heat produced by the disintegration of radioactive elements. However, both Holmes (1947, p. 123–126) and Jeffreys (1950, p. 119–120) believe the escape of internal heat contributes to the heat loss calculated at the surface. It is Holmes' belief that the sialic material contains 150 times as much uranium as average earth material. If that conclusion is correct, escaping internal heat should be considerably more important relatively in those vast parts of the globe that are not veneered by the thin sial.

Submarine Canyons and Allied Features

Factual Data

It was dissatisfaction with all current explanations for submarine canyons that started me on this study. The many theories advanced to explain these features are described and discussed by Shepard (1948, p. 238–244).

So much has been written about submarine canyons in recent years that it will not be necessary to describe them in detail. A most

comprehensive account of their distribution is given by Shepard (1948). They cut the submerged continental slopes and shelves of all continents and of many islands including the Bahamas (Ewing, Woollard, Vine, and Worzel, 1946, p. 925). Most of the larger canyons connect with rivers, and as a general rule with the larger rivers. Smaller canyons notch the continental slope but do not penetrate far into the continental shelf. Most of the canyons that do approach the present shore line have been diminished and even obliterated through infilling brought about by shoreline processes.

Many canyons have cut through the mantle into underlying bedrock. The bedrock encountered ranges in vulnerability to erosion from shale to granite. Studies made of the canyons off the California coast have shown the presence along the canyon walls of limestone, well cemented sandstone, conglomerate, basalt, and granite (Shepard and Emery, 1941, p. 88, 89) as well as soft Tertiary material. The maximum depth reached by the canyon floors is certainly greater than 12,000 feet and may even be greater than 15,000. This is the stumbling block that prevents widespread acceptance of canyon cutting by subaerial erosion.

The authorities are in rather general agreement that the canyon cutting took place during the Pleistocene. The bedrock incised is as young as late Tertiary (Stetson, 1936). Veatch and Smith (1939, p. 48) state that the greatest erosion along the Atlantic-facing continental slope was during the first and last glacial stages. Shepard (1931, p. 345) and others have pointed out that some of the canyons in higher latitudes were modified by glacial excavation. These canyons not only have the characteristic cross section of glaciated valleys, but some contain moraines.

Another submerged feature suggestive of erosion is the terrace. According to Tolstoy (1951, p. 441) "the continental slopes are now known to show in many cases a step-like succession of horizontal or imperceptibly sloping shelves or terraces". The same author notes evidence of terracing on the flanks of the flat-topped submerged mountains (seamounts or guyots) rising above the ocean floor southeast of Cape Cod. The mid-Atlantic Ridge between depths of 10,000 and 15,000 feet contains "a succession of flats which for the lack of any better term have been called terraces". Emery (1948, p. 857) describes "deep terrace-like structures" on the flanks of Eniwetok Atoll at a depth of about 4500 feet. This compares closely with the depth of the flat top of the seamount tied to the northwest side of Bikini Atoll by a narrow neck. Bikini itself has a terrace at a depth of 13,000 feet (Dobrin *et al.*, 1949, p. 808).

At least 35 mountains rise to heights of 3500 to 12,400 feet above the floor of the Gulf of Alaska (Menard and Dietz, 1951, p. 1263; Murray, 1941, p. 339). Some of these have the symmetry, the slope, and the alignment of volcanoes. Most of the shallower mountains have flat tops at a depth of 2400 to 3000 feet. Some of the flat tops are as much at 8 and 9 miles in diameter. Similar seamounts have been found elsewhere, especially in the western and central Pacific (Hess, 1946). These truncated cones rise 9000 to 12,000 feet above the ocean floor, and the flat summit levels range as a general rule from 3000 to 6000 feet below sea level. The flat top is in many instances bordered by a gently sloping shelf which extends outward 1–2 miles. The summit levels of adjacent peaks may differ by as much as 1000 feet. Many seamounts have been used as foundations for atoll construction.

There is fairly general agreement that seamounts are volcanic cones that have been truncated by wave erosion. The flat submerged banks may have similar origin.

Ripple marks have been photographed on the sea floor at depths as great as 4500 feet. Shallow-water faunas, in addition to the reef-building types previously mentioned, have been found at even greater depths. Examples range from Foraminifera to larger mollusks (Shepard and Emery, 1941, p. 141–142). Sand, gravel, and even cobbles have been dredged from the floors of submarine canyons far from shore and at depths of 12,000 feet or more (Shepard, 1951a, p. 61). Recent coring in the Atlantic, along the eastern continental slope of North America, has shown the presence of beds of sand interbedded between layers of clay referred to as abyssal in type. However, I am skeptical in regard to the precise dating given these ocean-bottom deposits, based on Foraminifera. If I read Phleger (1951, p. 80) correctly all the Pleistocene species

of Foraminifera are still living. There are cold- and warm-water types, and as the ice sheets advanced these types shifted south, later to return with the retreat of the ice. Therefore a "normal post-Wisconsin assemblage of Foraminifera known to be living in the region at the present time" (Ericson, Ewing, and Heezen, 1951, p. 963) could equally well be assigned to one of the pre-Wisconsin interglacial stages, or perhaps even to an intersubstage of the Wisconsin itself.

Subaerial Erosion Hypothesis

The subaerial hypothesis as an explanation for the submarine canyons is as old as the discovery of the canyons themselves. It is based on the striking parallelism between canyons cut by rivers running off the land and the canyons cut into the submerged continental slopes and shelves. Shepard (1948, p. 229–231) has summarized the evidence favoring subaerial erosion of the submarine canyons. These valleys resemble land canyons in every possible respect with one exception, which is discussed later. The submarine canyons have steep walls with V-shaped cross sections. They follow a winding course and have accordant tributaries. The dimensions are comparable to those of larger land canyons. Some have wide flat floors toward the outer end, and others terminate in deltas. Natural levees have been observed. Hess (1933) noted that some of the submerged river valleys of the Bahamas even show a trellis pattern, presumably due to the structural situation.

An undaunted few have tried to explain the extreme sea-level fluctuation necessary to permit canyon cutting by subaerial processes. Shepard (1948, p. 245–246; 1951b) and du Toit (1940) have invoked warping of the continental borders. Hess and MacClintock (1936), convinced that no answer other than subaerial erosion can be the right one, suggest as a possible solution a change in the ellipticity of the sea surface perhaps due to sudden decrease in rate of earth rotation. Von Engeln (1950, p. 162) postulates interior earth conditions which would produce a temporary increase in density. This would cause the ocean bottoms to sink to greater depths, and sea level would "decline in a significant degree".

My explanation is that during one of the Pleistocene glacial epochs the ocean basins subsided to such an extent that sea level was lowered the 15,000 feet or whatever is needed to account for subaerial canyon cutting. I have calculated that if all the sea floor now lying at depths below 13,300 feet (4000 meters) were to subside 20,000 feet, all the ocean floor above 13,300 feet of water depth would be emergent. This would expose all our submarine canyons except, perhaps, the mouths of the deepest ones. I cite this figure merely to show that unheard-of subsidences are not necessary to expose the canyon-cut continental slopes. Furthermore, this extreme subsidence only had to happen once. To obtain the other glacial epochs, a sea-level lowering of 3000 feet probably would have been sufficient. The deep-sea floor (below 4000 meters) would have to subside a little over 5000 feet to produce this emergence.

I do not know which of the four Pleistocene glacial epochs was the one of maximum withdrawal, but I suspect that it was one of the later ones, for there is some evidence of higher-level deltas through which the present canyons have cut. Each sea-level lowering would cause resumption of excavation and headward erosion in those parts of the valleys lying above that particular ocean-water level. Each resubmergence stopped the canyon cutting, but by the end of the Wisconsin epoch a few canyons had eroded headward across the entire width of the shelf. The upper ends of many of these submerged valleys were subsequently filled by shoreline processes of erosion and deposition.

Although a few others have been willing to consider a sea-level lowering of several thousands of feet, owing to ocean-floor subsidence, they have not been able to swallow a return of sea level to within a few hundreds of feet of its former position, and have therefore dropped the idea with considerable dispatch. According my concept of the mechanics of contraction, it would be even stranger if the sea level did not return to the same approximate level as before! When the ocean basins subsided, owing to a contracting interior, a condition of isostatic disequilibrium was established. Very shortly, in terms of geologic time, the continents also subsided and came to rest at the level of isostatic equilibrium.

At least two other arguments have been raised opposed to subaerial erosion of the under-water

canyons. The one exception to the complete parallelism between submerged canyons and those on the continents is the long profile (Woodford, 1951). In the case of the mature land streams this profile is concave. It is not concave in the Monterey sea valley and in other submarine canyons with the possible exception of the Congo. However, I do not believe that one should expect a young consequent stream running down the continental slope to show a normal profile. It was busily engaged in converting its initial slope to a normal profile when submergence interrupted the job. The mighty Congo came closer to finishing this conversion than most other streams. Another argument is the cutting of canyons in partially enclosed seas such as the Mediterranean (Kuenen, 1950, p. 509) and the Sea of Japan to depths far below the level of the sill separating the sea from the open ocean. The best explanation that comes to mind is that graben faulting within the enclosed sea has dropped the canyon area below its original level.

I concur with Hess and others in believing that the flat-topped seamounts or guyots are volcanoes truncated by wave erosion. The lowering of present sea level from 3000 to 6000 feet would be adequate. The discordance in summit elevations can be explained in two ways: (1) the higher volcanoes (in terms of depth to the flat top) were not even in existence when the lower ones were truncated by wave action, or (2) erratic grabening of the fragmented Pacific Ocean floor has dropped the seamounts varying distances.

Other Suggested Explanations

Opposed to the concept of subaerial erosion are some of the most highly respected and revered heroes of modern geology. Most of the substitute ideas produced by this galaxy of mental stars although ingenious are none the less unpalatable. The leading alternate explanation is the turbidity-current hypothesis. Sired by Daly (1942, p. 147) and nurtured by Kuenen (various papers, 1937 to 1951), and by Ewing and associates, the turbidity-current hypothesis reached maturity with the publication of a symposium in 1951 (Society of Economic Paleontologists and Mineralogists).

Briefly, turbidity currents are silty underflows that have been observed in Lake Mead and several other artificial lakes in the United States as well as Lakes Geneva and Constance in Europe. They have also been produced synthetically in a tank (Kuenen and Migliorini, 1950), and there is no doubt that under proper conditions silty water slides down the sloping floor of reservoir, lake, or tank much like water flows under air.

Proponents of turbidity currents credit them with depositing the deep offshore graded sand deposits (the coarser clastics such as the cobbles are supposed to ride down on submarine landslides) and with carrying shallow-water Foraminifera out to depths far beyond their normal habitat, as well as cutting canyons into the continental slope and shelf.

Turbidity currents have not been found in the ocean in spite of the fact that some mighty rivers carry sediment to the heads of submarine canyons having gradients far steeper than those found on the floor of Lake Mead. It is perhaps more than a coincidence that none of the witnessed examples of turbidity currents have been where silt-laden masses of fresh water entered larger masses of salt water. Where rivers, such as the Congo, enter the sea, the fresh water rides out over the salt water, carrying its sediment with it until the checking of velocity causes the sediment to sink into and through the underlying quiet salt water

I claim that the finding of graded clastics and misplaced (shallow-water) faunas deep beneath the sea is not *prima facie* evidence that they were carried there by turbidity currents; that the finding of cobbles does not prove that they were transported by submarine landslides; and that photographs of ripple marks lying at a depth of 4500 feet do not necessarily mean that they resulted from current action operating at that depth. In any other environment, sand, gravel, and cobbles are recognized as stream deposits, especially where they occur on canyon floors, and in the fans and deltas at the mouths of canyons. Misplaced faunas are found at great depths in rigid reefs, and their position there is generally recognized as due to subsidence, but we are told that loose, misplaced forams must have been carried in! I likewise believe that the deep sea-floor current ripples, like the truncated seamounts, are relics of shallower water.

What manner of logic allows us to accept

evidence, such as marine strata, of a sea-level rise above present datum of 25,000 feet, but causes us to run from evidence of a sea-level depression of 25,000 feet? By what reasoning do we assume that current sea level is also (within 200 or 300 feet) minimum sea level? Would it not be just as logical to assume that the present sea level (when the continents and the ocean basins are in isostatic adjustment) is closer to median sea level? And when we do lower sea level a couple of hundred feet why do we limit that lowering to the amount of water that could have been removed to make ice caps and mountain glaciers? We raise sea level thousands of feet by diastrophism. Is this a one-way process? Why can't we lower sea level the same way? What is so sacrosanct about current sea level?

Glacial and Abnormally Warm Climates

There have been at least four periods of glaciation. These were in the Quaternary, Upper Carboniferous, close of the Proterozoic, and in the earlier Precambrian. The glaciation at the close of the Proterozoic was "immediately preceding and perhaps extending into the Cambrian period" (Brooks, 1949, p. 177). Both Schuchert (1914, p. 286) and Brooks point to the close relationship between these periods of ice formation and the occurrence of intensive mountain building accompanied by elevation. Rastall (1929, p. 448) emphasizes the low-latitude occurrence of the pre-Pleistocene glaciations.

Beginning with Dana, a number of geologists have suggested uplifts of the continents or sinking of sea level as a cause of glaciation (Flint, 1947, p. 503–504; Enquist, 1915). Some of the earlier advocates of the elevation hypothesis, among them Upham (1893), cited the submarine canyons then known as further evidence of lowered sea level.

There can be no doubt that a lowering of sea level will be accompanied by a decrease in temperature. Brooks (1949, p. 212) estimates the average decrease in temperature with increase above sea to be 0.3 degree Fahrenheit for each 100 feet. In addition, the increased continentality, decreased ocean currents, and increased volcanicity that would accompany a lowering of sea level would tend further to increase the rate of temperature decline.

According to this concept of the earth-contraction hypothesis, the glacial climates were brought about by the subsidence of the ocean floors to such depths that sea level was lowered at least 3000 feet. We have already seen that at least once during the Pleistocene the sea level was lowered to a considerably greater figure. The glacial climates were terminated by the subsidence of the continental blocks to approximately the same relative levels as they had occupied previously, thereby restoring the isostatic adjustment.

The presence in the Pleistocene of at least four distinct glacial climates has been the leading deterrent to widespread adoption of the glacial climate-by-elevation hypothesis in recent years. According to the ideas expressed here, the oceanic segments sank four times in a row with the continental segments following them down each time, but not until adequate time had elapsed for a glacial epoch. This would involve an aggregate drop during the Pleistocene alone of at least 8 miles, counting 25,000 feet for the deep canyon-producing subsidence and 5000 feet for each of the other three glacial epochs. However, if the contraction by solidification thesis is accepted, we have up to 140 miles to spend in this way.

Abnormally warm (in terms of the present) climates were the result of the subsidence and widespread inundation of peneplained continental platforms.

Successive Submergences and Emergences

General Background

Our geologic history is built upon the sedimentation record produced by a continuing series of submergences of vast areas of the continental platforms beneath the sea. Furthermore, the fossil evidence is to the effect that many of the great submergences were concurrent the world over. No adequate hypothesis has been advanced to explain these countless widespread immersions. Likewise unexplained are the epeirogenic movements that lifted the continental platforms high above the sea so that not only did the waters of the ocean withdraw but also the streams draining the continents were rejuvenated and great mountain chains were carved from the more resistant rocks. Again, these movements cannot all be

assigned to different times at different places. Many of them were worldwide in scope. The best example is the latest of these epeirogenic uplifts, which began in the late Tertiary and which is responsible for the world-encircling mountain systems of today.

I believe the so-called epeirogenic uplifts to be actually the lagging behind of the sial-capped continents in the general subsidence of crustal segments owing to contraction over a shrinking interior. As stated before, this lag resulted in an isostatic disequilibrium and so was short-lived. When the continents followed the ocean basins down, restoring the isostatic balance, submergence occurred where one or more of three conditions prevailed: (1) a tilting of the continental segments (owing to rotation of the block) took place during subsidence; (2) the borders of the continental platform had been dragged down when the oceanic segments subsided; and (3) sagging occurred in the sial-capped block, owing to fragmentation and local subsidence. I believe the last to be of greatest importance; the downwarping of basins and geosynclines caused the flooding of wide areas. The extent of the immersion was a function of (1) the topographic relief, and (2) the depth of the submergence.

Bridges and Moats

With some exceptions the paleontologists and the biologists have insisted that at times during the geologic past the continents were connected by land "bridges", and conversely there were periods when currently connected land areas were isolated from each other by water-filled "moats". Schuchert (1932, p. 879) believes that the distribution of life on opposite sides of the South Atlantic throughout most of the Paleozoic and Mesozoic "is overwhelmingly in favor of the existence of the Gondwana land bridge". He believes that this bridge submerged during the Tertiary. Willis (1932) suggests that land bridges are upthrust mountain ranges.

Few doubt the existence of a land bridge across Bering Strait during various times in the geologic past. According to Hibbard (1951, p. 204):

> "This land bridge between Alaska and Asia in the region of the Bering Strait existed intermittently during geologic time. When it was present during the Pleistocene, it was broad and high enough above sea level to support a good grassland, shrubbery, and some trees along the grassland border. The mammals that came from Asia to North America or went from North America to Asia either grazed and browsed their way across or preyed upon other animals that lived in this area as they moved from one continent to the other by population spread."

According to my modification of the contraction hypothesis, a segment of the "megabreccia" either lagging behind or moving ahead of its neighbors would, if in the proper geographic position, function as a bridge or moat until further movements changed the picture. We view with complacency faulting with vertical displacements of 18,000 feet (as along the Wasatch front) in the continental areas, but look with displeasure upon past faults of 8,000 feet that would drop a land bridge connecting South America and Africa to the present level of the Atlantic floor!

Intervals

The geologic record contains many "lost week-ends"[1] during which considerable organic evolution took place of which there is no record. The most notable of these intervals is the Lipalian between the Proterozoic and the Paleozoic during which life progressed, insofar as the fossil record is concerned, from simple algae, sponges, and worms to such complex forms as trilobites. As a matter of fact, even the presence of the simple forms in the Precambrian lacks complete proof, for none of the alleged fossils so far discovered is beyond suspicion (Raymond, 1935). Furthermore, the absence of good fossils cannot be charged to the excessive metamorphism of these ancient rocks, for we now recognize thick sections of unmetamorphosed sediments as of probable Precambrian age.

Such advanced animals as trilobites must have been in existence for millions of years before "their first actual appearance as fossils in early Cambrian strata" according to Raymond (1935, p. 385–386). He believes that trilobites were actually present in the Precam-

[1] In case this address should ever be read by posterity the "lost week-end" refers to a recent book and moving picture (cinema) which I understand covers a period of time blotted out as far as the leading character is concerned because of excessive intake of alcoholic beverages.

brian but were possessed of no hard parts. There would still seem to be a considerable gap in the evolutionary record, however, unless one is willing to admit that the trilobites and the shell fish donned their overcoats overnight.

My suggestion is that a sinking of the oceanic basins withdrew the waters from the continents, and it was millions of years before those parts of the continents now above water were again inundated. It is also possible that there was less water in the oceans at this time, so the continents had to subside to unusual levels in order to become immersed. The time between the youngest Precambrian sample analyzed and the oldest Cambrian specimen is in the neighborhood of 140 million years according to measurements based on radioactivity (Holmes, 1947, p. 105). Even if the Lipalian interval occupies only half of this span it would still be longer than the intervals ascribed to most of the post-Proterozoic periods and longer than the combined span of the Tertiary and Quaternary periods. With that perspective the appearance of life in fairly advanced form in the Cambrian does not seem quite so "explosive."

The Lipalian is only one of many intervals of missing record. Major breaks in the evolutionary record occur between eras, periods, and even subperiods. No doubt during these intervals sedimentation continued and life existed and evolved, but the records of the time are buried in the deposits now lying beneath the waters of the oceans.

The importance of the sea withdrawals was not lost upon Ulrich (1911, p. 494):

> "... the almost total absence of gradation between the specific stages in the evolution of the genera proves that the transition from one species to the next in its line usually took place in times not represented by deposits in the local sedimentary record. As it is unreasonable to assume that, with continuous opportunity, only the finished specific or varietal stage entered the areas of accessible depositional record, and as the process of mutation, while not uniform in rate, is yet continuous, the conclusion is inevitable that long periods intervened in which deposition was interrupted by sea withdrawal and during which mutation progressed to some later distinguishable stage."

This thought can be carried still further. Perhaps the time intervals covered by the actual inundations of the continents represent only a small fraction of total geologic time. Isostatic equilibrium between continental platforms and ocean basins demands a relationship approximately similar to that of today with the continents largely emergent. If emergent continents were the rule rather than the exception, our available sedimentary sections and their contained fossils must be considered brief, irregularly spaced, and totally inadequate samples of the sedimentation and evolutionary history of the earth. Around the borders of the continental platforms, now submerged beneath deep ocean water, are the many missing chapters of earth history. Here lie buried treasures of infinitely greater interest than the gold coins and jewelled trinkets allegedly buried by Captain Kidd!

Conclusion

Most of the ideas I have presented here are not new; they have evolved in the minds of others, but standing alone they have not been generally accepted. My objective has been to knit these individual ideas into a whole that is more plausible than the sum of the plausibilities of its parts.

I close with a quoted admonition: "Our greatest handicap in our search for the ultimate truth is the inherent conservatism of our own minds."[2]

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[2] My apologies to someone. I read this sentence somewhere, memorized it, and then forgot who said it!

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canyons. The one exception to the complete parallelism between submerged canyons and those on the continents is the long profile (Woodford, 1951). In the case of the mature land streams this profile is concave. It is not concave in the Monterey sea valley and in other submarine canyons with the possible exception of the Congo. However, I do not believe that one should expect a young consequent stream running down the continental slope to show a normal profile. It was busily engaged in converting its initial slope to a normal profile when submergence interrupted the job. The mighty Congo came closer to finishing this conversion than most other streams. Another argument is the cutting of canyons in partially enclosed seas such as the Mediterranean (Kuenen, 1950, p. 509) and the Sea of Japan to depths far below the level of the sill separating the sea from the open ocean. The best explanation that comes to mind is that graben faulting within the enclosed sea has dropped the canyon area below its original level.

I concur with Hess and others in believing that the flat-topped seamounts or guyots are volcanoes truncated by wave erosion. The lowering of present sea level from 3000 to 6000 feet would be adequate. The discordance in summit elevations can be explained in two ways: (1) the higher volcanoes (in terms of depth to the flat top) were not even in existence when the lower ones were truncated by wave action, or (2) erratic grabening of the fragmented Pacific Ocean floor has dropped the seamounts varying distances.

Other Suggested Explanations

Opposed to the concept of subaerial erosion are some of the most highly respected and revered heroes of modern geology. Most of the substitute ideas produced by this galaxy of mental stars although ingenious are none the less unpalatable. The leading alternate explanation is the turbidity-current hypothesis. Sired by Daly (1942, p. 147) and nurtured by Kuenen (various papers, 1937 to 1951), and by Ewing and associates, the turbidity-current hypothesis reached maturity with the publication of a symposium in 1951 (Society of Economic Paleontologists and Mineralogists).

Briefly, turbidity currents are silty underflows that have been observed in Lake Mead and several other artificial lakes in the United States as well as Lakes Geneva and Constance in Europe. They have also been produced synthetically in a tank (Kuenen and Migliorini, 1950), and there is no doubt that under proper conditions silty water slides down the sloping floor of reservoir, lake, or tank much like water flows under air.

Proponents of turbidity currents credit them with depositing the deep offshore graded sand deposits (the coarser clastics such as the cobbles are supposed to ride down on submarine landslides) and with carrying shallow-water Foraminifera out to depths far beyond their normal habitat, as well as cutting canyons into the continental slope and shelf.

Turbidity currents have not been found in the ocean in spite of the fact that some mighty rivers carry sediment to the heads of submarine canyons having gradients far steeper than those found on the floor of Lake Mead. It is perhaps more than a coincidence that none of the witnessed examples of turbidity currents have been where silt-laden masses of fresh water entered larger masses of salt water. Where rivers, such as the Congo, enter the sea, the fresh water rides out over the salt water, carrying its sediment with it until the checking of velocity causes the sediment to sink into and through the underlying quiet salt water

I claim that the finding of graded clastics and misplaced (shallow-water) faunas deep beneath the sea is not *prima facie* evidence that they were carried there by turbidity currents; that the finding of cobbles does not prove that they were transported by submarine landslides; and that photographs of ripple marks lying at a depth of 4500 feet do not necessarily mean that they resulted from current action operating at that depth. In any other environment, sand, gravel, and cobbles are recognized as stream deposits, especially where they occur on canyon floors, and in the fans and deltas at the mouths of canyons. Misplaced faunas are found at great depths in rigid reefs, and their position there is generally recognized as due to subsidence, but we are told that loose, misplaced forams must have been carried in! I likewise believe that the deep sea-floor current ripples, like the truncated seamounts, are relics of shallower water.

What manner of logic allows us to accept

evidence, such as marine strata, of a sea-level rise above present datum of 25,000 feet, but causes us to run from evidence of a sea-level depression of 25,000 feet? By what reasoning do we assume that current sea level is also (within 200 or 300 feet) minimum sea level? Would it not be just as logical to assume that the present sea level (when the continents and the ocean basins are in isostatic adjustment) is closer to median sea level? And when we do lower sea level a couple of hundred feet why do we limit that lowering to the amount of water that could have been removed to make ice caps and mountain glaciers? We raise sea level thousands of feet by diastrophism. Is this a one-way process? Why can't we lower sea level the same way? What is so sacrosanct about current sea level?

Glacial and Abnormally Warm Climates

There have been at least four periods of glaciation. These were in the Quaternary, Upper Carboniferous, close of the Proterozoic, and in the earlier Precambrian. The glaciation at the close of the Proterozoic was "immediately preceding and perhaps extending into the Cambrian period" (Brooks, 1949, p. 177). Both Schuchert (1914, p. 286) and Brooks point to the close relationship between these periods of ice formation and the occurrence of intensive mountain building accompanied by elevation. Rastall (1929, p. 448) emphasizes the low-latitude occurrence of the pre-Pleistocene glaciations.

Beginning with Dana, a number of geologists have suggested uplifts of the continents or sinking of sea level as a cause of glaciation (Flint, 1947, p. 503–504; Enquist, 1915). Some of the earlier advocates of the elevation hypothesis, among them Upham (1893), cited the submarine canyons then known as further evidence of lowered sea level.

There can be no doubt that a lowering of sea level will be accompanied by a decrease in temperature. Brooks (1949, p. 212) estimates the average decrease in temperature with increase above sea to be 0.3 degree Fahrenheit for each 100 feet. In addition, the increased continentality, decreased ocean currents, and increased volcanicity that would accompany a lowering of sea level would tend further to increase the rate of temperature decline.

According to this concept of the earth-contraction hypothesis, the glacial climates were brought about by the subsidence of the ocean floors to such depths that sea level was lowered at least 3000 feet. We have already seen that at least once during the Pleistocene the sea level was lowered to a considerably greater figure. The glacial climates were terminated by the subsidence of the continental blocks to approximately the same relative levels as they had occupied previously, thereby restoring the isostatic adjustment.

The presence in the Pleistocene of at least four distinct glacial climates has been the leading deterrent to widespread adoption of the glacial climate-by-elevation hypothesis in recent years. According to the ideas expressed here, the oceanic segments sank four times in a row with the continental segments following them down each time, but not until adequate time had elapsed for a glacial epoch. This would involve an aggregate drop during the Pleistocene alone of at least 8 miles, counting 25,000 feet for the deep canyon-producing subsidence and 5000 feet for each of the other three glacial epochs. However, if the contraction by solidification thesis is accepted, we have up to 140 miles to spend in this way.

Abnormally warm (in terms of the present) climates were the result of the subsidence and widespread inundation of peneplained continental platforms.

Successive Submergences and Emergences

General Background

Our geologic history is built upon the sedimentation record produced by a continuing series of submergences of vast areas of the continental platforms beneath the sea. Furthermore, the fossil evidence is to the effect that many of the great submergences were concurrent the world over. No adequate hypothesis has been advanced to explain these countless widespread immersions. Likewise unexplained are the epeirogenic movements that lifted the continental platforms high above the sea so that not only did the waters of the ocean withdraw but also the streams draining the continents were rejuvenated and great mountain chains were carved from the more resistant rocks. Again, these movements cannot all be

assigned to different times at different places. Many of them were worldwide in scope. The best example is the latest of these epeirogenic uplifts, which began in the late Tertiary and which is responsible for the world-encircling mountain systems of today.

I believe the so-called epeirogenic uplifts to be actually the lagging behind of the sial-capped continents in the general subsidence of crustal segments owing to contraction over a shrinking interior. As stated before, this lag resulted in an isostatic disequilibrium and so was short-lived. When the continents followed the ocean basins down, restoring the isostatic balance, submergence occurred where one or more of three conditions prevailed: (1) a tilting of the continental segments (owing to rotation of the block) took place during subsidence; (2) the borders of the continental platform had been dragged down when the oceanic segments subsided; and (3) sagging occurred in the sial-capped block, owing to fragmentation and local subsidence. I believe the last to be of greatest importance; the downwarping of basins and geosynclines caused the flooding of wide areas. The extent of the immersion was a function of (1) the topographic relief, and (2) the depth of the submergence.

Bridges and Moats

With some exceptions the paleontologists and the biologists have insisted that at times during the geologic past the continents were connected by land "bridges", and conversely there were periods when currently connected land areas were isolated from each other by water-filled "moats". Schuchert (1932, p. 879) believes that the distribution of life on opposite sides of the South Atlantic throughout most of the Paleozoic and Mesozoic "is overwhelmingly in favor of the existence of the Gondwana land bridge". He believes that this bridge submerged during the Tertiary. Willis (1932) suggests that land bridges are upthrust mountain ranges.

Few doubt the existence of a land bridge across Bering Strait during various times in the geologic past. According to Hibbard (1951, p. 204):

> "This land bridge between Alaska and Asia in the region of the Bering Strait existed intermittently during geologic time. When it was present during the Pleistocene, it was broad and high enough above sea level to support a good grassland, shrubbery, and some trees along the grassland border. The mammals that came from Asia to North America or went from North America to Asia either grazed and browsed their way across or preyed upon other animals that lived in this area as they moved from one continent to the other by population spread."

According to my modification of the contraction hypothesis, a segment of the "megabreccia" either lagging behind or moving ahead of its neighbors would, if in the proper geographic position, function as a bridge or moat until further movements changed the picture. We view with complacency faulting with vertical displacements of 18,000 feet (as along the Wasatch front) in the continental areas, but look with displeasure upon past faults of 8,000 feet that would drop a land bridge connecting South America and Africa to the present level of the Atlantic floor!

Intervals

The geologic record contains many "lost week-ends"[1] during which considerable organic evolution took place of which there is no record. The most notable of these intervals is the Lipalian between the Proterozoic and the Paleozoic during which life progressed, insofar as the fossil record is concerned, from simple algae, sponges, and worms to such complex forms as trilobites. As a matter of fact, even the presence of the simple forms in the Precambrian lacks complete proof, for none of the alleged fossils so far discovered is beyond suspicion (Raymond, 1935). Furthermore, the absence of good fossils cannot be charged to the excessive metamorphism of these ancient rocks, for we now recognize thick sections of unmetamorphosed sediments as of probable Precambrian age.

Such advanced animals as trilobites must have been in existence for millions of years before "their first actual appearance as fossils in early Cambrian strata" according to Raymond (1935, p. 385–386). He believes that trilobites were actually present in the Precam-

[1] In case this address should ever be read by posterity the "lost week-end" refers to a recent book and moving picture (cinema) which I understand covers a period of time blotted out as far as the leading character is concerned because of excessive intake of alcoholic beverages.

brian but were possessed of no hard parts. There would still seem to be a considerable gap in the evolutionary record, however, unless one is willing to admit that the trilobites and the shell fish donned their overcoats overnight.

My suggestion is that a sinking of the oceanic basins withdrew the waters from the continents, and it was millions of years before those parts of the continents now above water were again inundated. It is also possible that there was less water in the oceans at this time, so the continents had to subside to unusual levels in order to become immersed. The time between the youngest Precambrian sample analyzed and the oldest Cambrian specimen is in the neighborhood of 140 million years according to measurements based on radioactivity (Holmes, 1947, p. 105). Even if the Lipalian interval occupies only half of this span it would still be longer than the intervals ascribed to most of the post-Proterozoic periods and longer than the combined span of the Tertiary and Quaternary periods. With that perspective the appearance of life in fairly advanced form in the Cambrian does not seem quite so "explosive."

The Lipalian is only one of many intervals of missing record. Major breaks in the evolutionary record occur between eras, periods, and even subperiods. No doubt during these intervals sedimentation continued and life existed and evolved, but the records of the time are buried in the deposits now lying beneath the waters of the oceans.

The importance of the sea withdrawals was not lost upon Ulrich (1911, p. 494):

> "... the almost total absence of gradation between the specific stages in the evolution of the genera proves that the transition from one species to the next in its line usually took place in times not represented by deposits in the local sedimentary record. As it is unreasonable to assume that, with continuous opportunity, only the finished specific or varietal stage entered the areas of accessible depositional record, and as the process of mutation, while not uniform in rate, is yet continuous, the conclusion is inevitable that long periods intervened in which deposition was interrupted by sea withdrawal and during which mutation progressed to some later distinguishable stage."

This thought can be carried still further. Perhaps the time intervals covered by the actual inundations of the continents represent only a small fraction of total geologic time. Isostatic equilibrium between continental platforms and ocean basins demands a relationship approximately similar to that of today with the continents largely emergent. If emergent continents were the rule rather than the exception, our available sedimentary sections and their contained fossils must be considered brief, irregularly spaced, and totally inadequate samples of the sedimentation and evolutionary history of the earth. Around the borders of the continental platforms, now submerged beneath deep ocean water, are the many missing chapters of earth history. Here lie buried treasures of infinitely greater interest than the gold coins and jewelled trinkets allegedly buried by Captain Kidd!

Conclusion

Most of the ideas I have presented here are not new; they have evolved in the minds of others, but standing alone they have not been generally accepted. My objective has been to knit these individual ideas into a whole that is more plausible than the sum of the plausibilities of its parts.

I close with a quoted admonition: "Our greatest handicap in our search for the ultimate truth is the inherent conservatism of our own minds."[2]

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[2] My apologies to someone. I read this sentence somewhere, memorized it, and then forgot who said it!

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PALYNOLOGY IN SOUTHERN NORTH AMERICA. I: ARCHEOLOGICAL HORIZONS IN THE BASINS OF MEXICO

BY PAUL B. SEARS

ABSTRACT

Profiles of sediments in the Basin of Mexico show an inverse relation between the percentages of fossil pine and oak pollen. Vegetation studies suggest that any marked increase in oak indicates an increase in available moisture, while consistently low percentages of oak indicate moisture deficiency. On this basis prolonged moist and dry periods have alternated. The more recent of these climatic shifts can be correlated stratigraphically with known phases of human occupation (Sears, 1951b).

The Archaic Culture, 2000 or 1500 to 500 or 400 B.C., and the Nahua, 800 or 900 to 1521 A.D., both flourished within the Basin during relatively moist conditions. These two humid periods were separated by a prolonged dry interval which overtook the Late Archaic during a time of very low lake level.

The Late Archaic level was then covered by a volcanic ash fall, and the culture shifted northward to higher ground. There it became the Teotihuacán which lasted from 500 or 400 B.C. to 800 or 900 A.D. Since this highly developed culture existed during a dry period, it must have made use of ground water. If so, the sources should have been unfavorably affected by heavy soil erosion known to have occurred, and by any deforestation, as Vaillánt (1941) suggests in explanation of the ultimate Teotihuacán collapse.

Conclusions based on pollen analysis are consistent with what is known of soil profiles and former lake levels.

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Introduction

Pollen analysis or palynology has been chiefly used in tracing climatic change, although its value in stratigraphic correlation is increasingly recognized. Most of the sediments employed for climatic studies have been within glaciated areas, hence their record applies only to times when the basins containing them have been ice-free.

The basins considered in the present series of studies lie 1500–3000 international air km beyond the border of continental Pleistocene glaciation. Fossil pollen has already been recovered from them at depths of more than 60 meters, giving a continuous record extending back into the Pleistocene. Such a record has the added value of showing the degree to which climatic changes associated with known advances and retreats of the ice were effective at these distances beyond the glacial limits (Sears, 1951a).

This initial paper, however, deals only with the analysis of relatively shallow sediments which, in the Basin of Mexico, can be correlated with certain known phases of human culture in that area.

The principle involved is presented in Deevey's (1944) pioneer paper on pollen analysis in Mexico. Samples taken at suitable intervals in the sedimentary column are treated so as to free and concentrate pollen and other spores and prepare them for microscopic examination. Counts are then made, and percentages calculated for forest-tree pollen; all other forms are reckoned as so many per hundred of forest-tree pollen.

By this method major changes in forest composition are revealed, and with them significant changes in climate. The non-forest pollen in turn is useful not only for confirmation as to climate, but as an index to local changes, such as vulcanism, cultivation, or shifting water levels.

While Deevey's (1944) study was frankly exploratory, it revealed certain essential facts, here amply confirmed. Upland forest pollen was represented by fir, pine, alder, and oak, herbaceous flora by grasses, chenopods, and composites. Evidence of agriculture was afforded by the finding of a few grains of maize and agave pollen. A dry period within recent, but probably pre-historic, times was postulated.

Acknowledgments

In 1948 Dr. Pablo Martinez del Rio invited my attention to the problem of past climates in relation to human activity within the Basin of Mexico and has most generously assisted me throughout.

For invaluable guidance in the field, I am indebted to Ing. A. R. V. Arellano; Drs. Faustino Miranda, Manuel Maldonado-Koerdell, and Helmut de Terra; Senores Arturo Romano and Luis Aveleyra Arroya de Anda; Senora Antonieta Espejo, and Senor Pina-Chan. For generous provision of facilities, acknowledgments are due to Dr. Rubin de Borbolla, Director of the Museo Nacional, to the Escuela de Agricultura at Chapingo, and to Dr. Sanchez Marroquen of the Instituto Politecnico.

Many other individuals have, in one way or another, given help and encouragement. Aside from those named in table captions, I must mention Mr. Philip Clisby, Professor Victor Lytle, Mrs. Jean Galloway McNab, and the late Mr. Ralph Smith, a North American long resident in Mexico City who first informed me of the late Archaic site, El Tepalcate, which has provided invaluable stratigraphic evidence.

Acknowledgments are also due The Geological Society of America, The Viking (Wenner-Gren) Foundation, and the United Fruit Company for generous aid, and to Oberlin College for a leave of absence in 1948 as well as for laboratory space.

Statement of Problem

Two questions are involved in the present paper. Is there evidence of climatic change within the Basin of Mexico during the time of known human occupancy? If so, can such change be correlated with known cultural horizons? Both questions seem answerable in the affirmative.

The Environment

The general setting is given by the map (Fig. 1), kindly furnished, along with the levels here used, by Ing. Arellano. Essentially

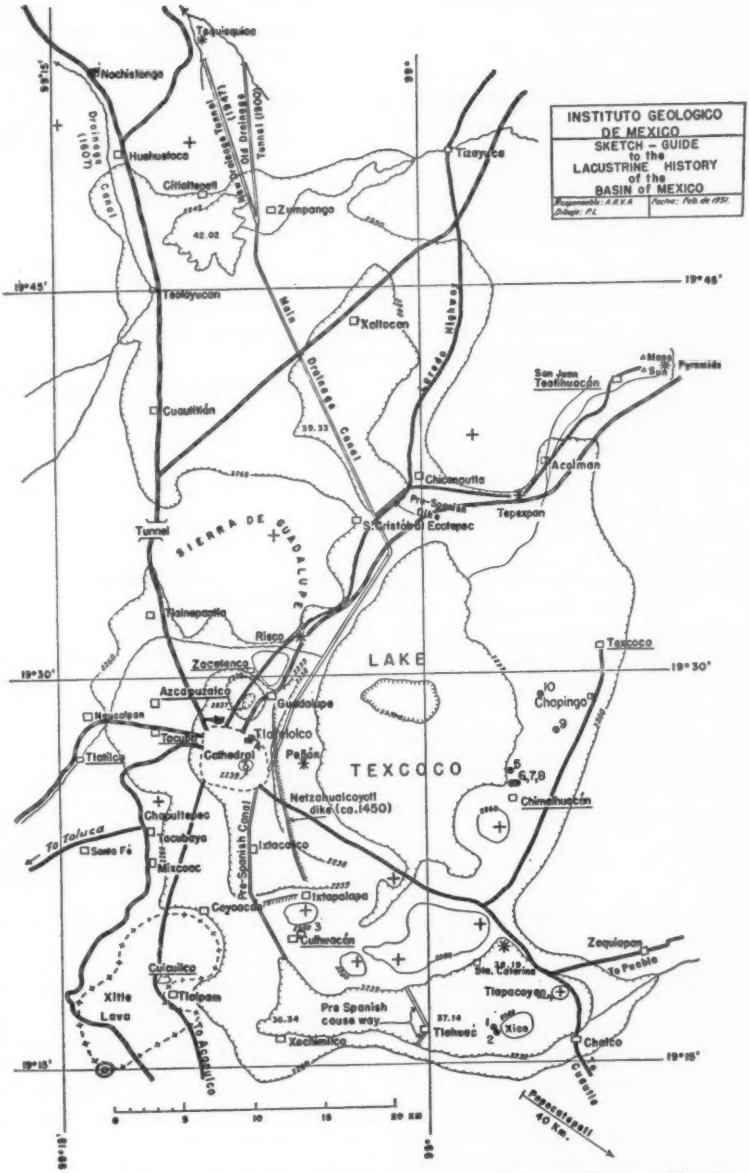

FIGURE 1.—SKETCH GUIDE TO THE LACUSTRINE HISTORY OF THE BASIN OF MEXICO

the map shows a large basin—not a valley—enclosed by mountains and displaying the evidence of a long volcanic and lacustrine history. From the remnant of Lake Texcoco on the present basin floor at 2235–2236 meters, traces of ancient beaches occur up to an elevation of

2263 meters, while sediments occur under the present City of Mexico to a depth of several hundred meters.

Since this is true, and there is a striking inverse relation between oak and pine in our data, we have assumed that oak indicates

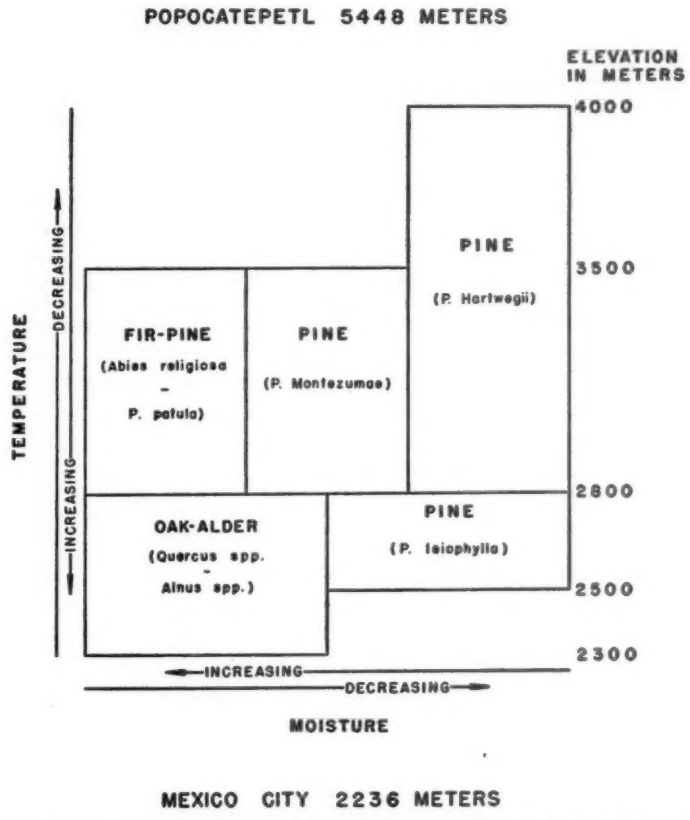

FIGURE 2.—INDICATOR SIGNIFICANCE OF THE FOUR PRINCIPAL UPLAND FOREST GENERA OF THE MEXICO CITY AREA

The highlands surrounding the basin are normally forested at levels of 2300 to 4000 meters. This forest consists principally of four genera—Fir (*Abies*), Pine (*Pinus*), Alder (*Alnus*), and Oak (*Quercus*)—whose general relation to altitude and moisture is shown in Figure 2. A number of species of both oak and pine are involved, not all in each genus having exactly the same requirements. Yet where oak and pine occur at the same altitude, oak is usually confined to ravines or areas having favorable soil moisture, while pine occupies the more exposed ridges, upland, and areas of dryer soil.

relatively humid, pine relatively dry conditions. This assumption has yielded results consistent with water levels, soil profiles, and frequency of nonforest pollen. Alder, which occurs along stream banks and as an understory in the stabilized forest, shows a rough correlation with oak as an indicator of relatively moist conditions.

At higher, colder altitudes, fir unquestionably occupies the more humid sites, although it shares them with one species of pine, *P. patula*. The dryer sites are occupied by other species of pine, so that any consistent increase of fir seems a safe index of increased moisture. Its

pollen grains are bulky and must come a considerable distance to be deposited in the basin, but I am not yet satisfied that our techniques are adequate to recover fir pollen in representative quantity. At any rate, it is not sufficiently abundant in the shallow layers here considered to affect conclusions.

Marked pulsations of climate within the valley are indicated by the existence of arid-climate caliches alternating with dark organic soils (Bryan, 1948; Arellano, 1951). There is also evidence of a series of glacial recessions and readvances on the slopes of Iztaccihuatl (De Terra, 1949). Lake levels within the basin afford a general confirmation, although purely physiographic factors must be allowed for here.

Archeology and History

1. Pre-Archaic Man

The presence of man before the extinction of the elephant fauna is supported by two skeletal finds, Peñon Man and Tepexpan Man, and by artifacts at several sites. For a critical evaluation of primitive artifacts regarded as preceding (San Juan), accompanying (Tepexpan), and following (Chalco) Tepexpan Man the reader may consult Aveleyra (1950).

Having examined both Peñon and Tepexpan sites, as well as the Peñon fossil, I do not believe these two finds can be written off lightly. The Tepexpan remains were certainly found in an elephant-bearing stratum, below an unconformity. They lay, moreover, in a zone subject to wind deflation during periods of low water level and must have been, at discovery, nearer the surface than for a long time past.

Pollen checks of Tepexpan material have been negative thus far, although new techniques will doubtless reveal pollen. Unfortunately, this could merely confirm the unquestioned antiquity of the stratum, without silencing certain critics who suggest that the find represents an intrusive burial.

2. Archaic or Middle Culture

This highly developed ceramic culture, whose derivation is unknown, flourished from about 2000 or 1500 B.C. to about 500 or 400 B.C., when its latest phase shifted northward to become Teotihuacán I. These and other dates are based upon the judgment of Mexican archeologists as generally confirmed by the radiocarbon studies of Arnold and Libby (1951).

The Early and Middle Archaic culture occupied upland sites, such as the high beach level of Zacatenco, whose elevation is about 2242 meters as compared with 2236, the present level of Mexico City. Water levels eventually dropped so much that the Late Archaic site of El Tepalcate was occupied at about 2235–2236 meters.

Important Archaic sites are shown in Figure 1. The rich classical site at Tlatilco occurs in alluvium, not favorable to pollen preservation and thus far negative. Other Archaic sites are found in the vicinity of the Pedregal lava flow which originated in Xitle. These include the spiral (caracol or snail) Archaic temple of Cuicilco, whose base was buried in lava, as well as the cultural layer now excavated from below the lava at Copilco.

Copilco shows plainly the presence of a dark organic soil which had been cultivated, and which must have been formed during a relatively humid period. Conditions at none of these sites are favorable for pollen stratigraphy.

Of the utmost importance in the present study is the Late Archaic site known as El Tepalcate, 2 kilometers north of Chimalhuacán. This site, which was under water at the time of the Conquest, had been occupied by Late Archaic at a low level of Lake Texcoco and has been exposed again by the recent artificial drainage of the Basin. Described by Apenes (1943), the site was excavated by Noguera (1943), who identified its cultural material with Teotihuacán I which he had found at the base of the pyramid of the Sun at Teotihuacán, thus establishing the direct link between that culture and the Archaic. He also noted that the cultural horizon at El Tepalcate was covered by a thick layer of sand. This "sand," as will be seen, is doubtless volcanic in origin, forming a deposit of considerable extent and marking the end of the Late Archaic at this site.

3. Teotihuacán Culture

Accepting Noguera's judgment, the Late Archaic now became Teotihuacán, situated at the relatively high level of about 2266 meters,

some 30 kilometers north and a little east of El Tepalcate, which had evidently become untenable. This transfer to higher ground at a time of extremely low water level within the basin raises an interesting question as to water supply, which will be discussed.

The center at Teotihuacán is marked by two well-known pyramids; the larger exceeds that of Cheops in volume and nearly equals it in height. These pyramids were flanked by extensive developments, believed to mark a great religious center. Lavish use was made of plaster, a fact which led Vaillant (1941) to attribute the abandonment of Teotihuacán (circa 800–900 A.D.) to the deforestation of the adjacent hills, on the assumption that vast amounts of fuel were used for the burning of lime. Cook (1949) reports evidence of serious erosion coincident with great population pressure here, a circumstance which lends support to Vaillant's idea of economic deterioration.

4. *Nahua Culture*

The Nahua, which includes the Aztec group, became established about 900. A.D. on promontories (Culhuacán, Tenochtitlán, Tlatelolco, etc.) within the Basin at a time of restored water level (circa 2239 meters). Moisture conditions were then favorable to the formation of dark organic soil, providing a rich food economy through horticulture, fishing, and hunting.

The system of gardening utilized *chinampas*, or beds formed by excavating canals at the shallow lake margins and piling the material above water line. The canals furnished irrigation and transport by dug-out canoe, while the sediment dredged from the lake was rich in mineral and organic material. The fisheries utilized a considerable variety of aquatic organisms. In addition there are still traces of a well-developed upland agriculture, now largely destroyed. The terraces were retained by maguey, a highly useful source of fiber and drink and an effective plant for erosion control, since it does not complete with maize. These remnants suggest that the Nahua conserved soil and moisture for upland production to supplement that of the *chinampas*.

Thus fortified by abundant food supplies, it is little wonder that the Nahua developed enough vitality to become overlords of Mexico within some four centuries.

5. *European Occupation*

Cook's studies show that indigenous cultures had been responsible for serious erosion toward the north of the Basin. Under the Nahua, however, forests appear to have been well conserved. With the Spanish Conquest of 1521 the role of man as a geological agent was stepped up, owing to the introduction of steel tools and domestic animals, and to the general emphasis on mining rather than agriculture. Unlike Costa Rica, which offered no mineral wealth, Mexico was not colonized primarily by skilled farmers. Despite some excellent examples of irrigation and contouring, heavy demands were made upon soil, water, and forests, while the hills were plowed and overrun by grazing animals.

The able and learned cosmographer, Henrico Martínez (1606), noting that plow, horse, and cattle were introduced by Christians, attributes the washing down of mud and earth into the valley thereafter to this fact. He describes the attendant stripping (*descarnado*) of the highlands down to *tepetate*, and specifically points out the resulting threat to the City of Mexico.

For reasons of political and religious policy, the Spaniards had rebuilt this city on the ruins of the old Nahua (Aztec) shrine and citadel, instead of on the safer and more salubrious hills. Actually the city had been flooded in Nahua times, first in 1446 or 1449, later in 1499 and 1517. Figure 1 shows a pre-Spanish dike built in consequence. A generation after the Conquest severe floods began to come at frequent intervals—1555, 1556, 1580, 1604, 1607 (Junta directiva, 1902). These resulted not in any project for removal to higher ground, but in a demand for drainage of the lakes, although these were the basis of the local food economy.

Under Martínez' direction a stupendous cut, the Tajo de Nochistongo, was completed by 1609, draining Lake Zumpango, at 2243 meters. But the lower lakes remained, and another disastrous flood came in 1629. Thorough drainage was not established until 1900. Thereafter

the bed of Lake Chalco became dry, and the level of Lake Texcoco was lowered to about 2235 meters, thus exposing the Late Archaic site, El Tepalcate. Even so, excessive rains can still produce floods. An aerial photograph taken about 1941 shows the bed of Texcoco filled almost to its old limits.

While the great drainage project now largely protects the city from flooding and affords space for industrial and residential purposes, it has greatly augmented the dust storms which during dry seasons are an almost daily feature within the old lake beds. I counted 23 at one time in the bed of Lake Chalco. Presently the individual storms coalesced, filling the entire atmosphere with dust. Drainage has also drastically reduced the harvest from lake fisheries and chinampa horticulture. The economy of such villages as Xico has been wrecked. Even the famous "floating gardens" or chinampas of Lake Xochimilco have suffered severely, despite efforts to maintain the water level while drawing upon it for public use (Anon., 1949). Water scarcity is today an obvious limiting factor within the Basin.

Whether this man-made aridity has been intensified by that secular climatic change which has resulted in general glacial retreat since about 1750 is not clear. Even farther north, where the facts of this retreat have been established, and where a general rise of temperature (resulting in lower relative humidity) seems to have taken place, problems of precise comparative measurement give rise to some uncertainty. So far as the pollen records go, deforestation of the past century makes interpretation of the most recent climate trends even more difficult.

Résumé

Evidence of marked climatic pulsations is afforded by alternating bands of mineral and organic soil, by moraine patterns at high elevation, and by changes in water level, although the latter must be used with caution.

Man's presence in the Basin with the Pleistocene fauna seems probable but has not yet been proved to general satisfaction.

The Archaic culture, beginning about 2000–1500 B. C., at first occupied relatively high beach levels and utilized a dark organic soil. This culture ended within the Basin at a time of minimum water level and considerable volcanic activity. It then moved northward, established itself on higher ground, and became Teotihuacán I. There it must have had some water source independent of the then low lakes, circa 500–400 B. C.

This impressive Teotihuacán culture, which ended circa 800–900 A. D. was responsible for heavy soil erosion and perhaps considerable deforestation. Whether the site became economically untenable is a question.

Nahua culture occupied the Basin, circa 900 A. D., when water levels were high enough to submerge the Late Archaic site on Lake Texcoco. Nahua culture was based largely on skillful use of aquatic resources.

Deforestation, overgrazing, and disastrous floods followed Spanish occupation in 1521. Drainage, undertaken to prevent floods, was finally accomplished about 1900, largely destroying the water economy which had sustained the Nahua and doubtless the Early and Middle Archaic.

Field Work

1948

Preliminary checks having shown the presence of pollen in peat from the Chalco-Xochimilco area, the dry lake margin near the hill, Xico, was selected for study. The village at the base of this hill had maintained an active commerce by canoe with Tenochtitlán in Mexica time and until the completion of drainage about 1900. The lake bed north and east of the rocky former island is dotted with low mounds or monticulos and strewn with potsherds and other archeological material, while a large hacienda, ruined during the recent revolution, attests the former productiveness of this now desolate and desiccated community.

Trenches were cut to determine the depth and character of human materials, and sampling was extended below trench bottom by posthole auger. One low mound was rather thoroughly explored, revealing a number of burials, wooden stakes at a depth of 1.8 meters, and an approach from the East. The most pertinent

discovery, for our purposes, was the clearly Nahua character of all artifacts, except a few Teotihuacán fragments judged by Sr. Aveleyra to be probably secondary, brought in by the Nahua occupants. This mound, 16–18 meters in diameter, rose about 1.6 meter above the general lake bed. Signs of disturbance, extending here and elsewhere to nearly 2 meters below the present lake bed, suggest that occupation was accompanied by rising water levels, an assumption borne out by the date of the first recorded flooding of Mexico City, 1446, or 1449.

White layers judged to be ash and containing pumice lay at variable depths, below the zone of occupation, but here and there had been disturbed by digging and intermingled with darker sediments. Analyses were made at two sites (Profiles 1 and 2).

Two other Nahua sites were examined. Culhuacán, between Mexico City and Xochimilco, is notable for its close and continuous series of stratified artifacts of Nahua origin. Below them is a thin floor of fine white diatomaceous sediment, with one or more heavier beds of volcanic ash interrupting the lake sediments below it. Santiago Tlatelolco, north of the cathedral which marks the old center of Mexico City, and formerly connected with it by causeway, is the site of a pyramid group. Excavation here has revealed eight, perhaps nine structures, each enclosing its predecessors as the latter sank and were partially buried by sediment. Collections were made below the Nahua zone at Culhuacán, and below the base of the oldest pyramid at Tlatelolco.

Three Archaic sites near Mexico City were also studied—the high Early Archaic beach at Zacatenco, north of the City, the rich alluvial site of Copilco in the valley of the Rio Hondo not far from Los Remedios, and the dark agricultural soil now buried beneath the Pedregal lava, south of the City. No pollen was found, probably due to oxidation and soil-forming processes.

Finally, and most fortunately, the Late Archaic site of El Tepalcate, near the present margin of Lake Texcoco, and about 2 kilometers north of Chimalhuacán, was excavated. A dark organic layer below the artifacts was sampled for analysis. In addition, although I had not, at the time, seen the paper by Noguera (1943), the overlying sand attracted my attention. Under the microscope the sand showed numerous shards and was in my judgment, later confirmed by Doctor de Terra, a volcanic deposit, *in situ*.

1949

Since any volcanic deposit in known relation to a culture horizon should afford a stratigraphic key, attention during this collecting season was concentrated on the Chapingo-Chimalhuacán area along the eastern edge of the bed of Lake Texcoco, centering about El Tepalcate.

Sampling was greatly facilitated by the use of aluminum alloy extensions for the Swedish Hiller borer, with spring-pin and sleeve connections (Fig. 3). This had been devised and made at Oberlin (Lichtwardt, MS) and was subjected to its first rigorous field test at this time. As before, the untiring help of Sr. Romano proved indispensable. A line of pits extending southward from El Tepalcate toward Chimalhuacán was dug, confirming the fact that the coarse, sandy, volcanic layer which overlies the Late Archaic is continuous, generally appearing as a hard-pan covered by more recent sediments. A spot was then selected in wet meadow 1 km north of Chimalhuacán where, according to our information, a lake had persisted until about 20 years ago. Here the solidified, sandy layer was encountered at a depth of somewhat over 2 meters (Fig. 4, Profiles 6 and 7). One set of samples was taken to a depth of 8.25 meters (Profile 8), thus giving a record of conditions preceding and following the deposition of the layer which apparently marks the end of the Late Archaic.

Another core was taken to a depth of 2.25 m in the rush turf bordering the present lake, just north of the Archaic site and hence equivalent except for the absence of artifacts. Here the sandy layer was at a depth of 75–100 cms and was slightly covered with recent organic material and turf (Profile 5).

A drilling was also made within the dry margin of Lake Texcoco, about 3.75 kms southwest of Chapingo (Profile 9). Here the same coarse, solidified sandy deposit, highly saline, has been exposed by wind deflation.

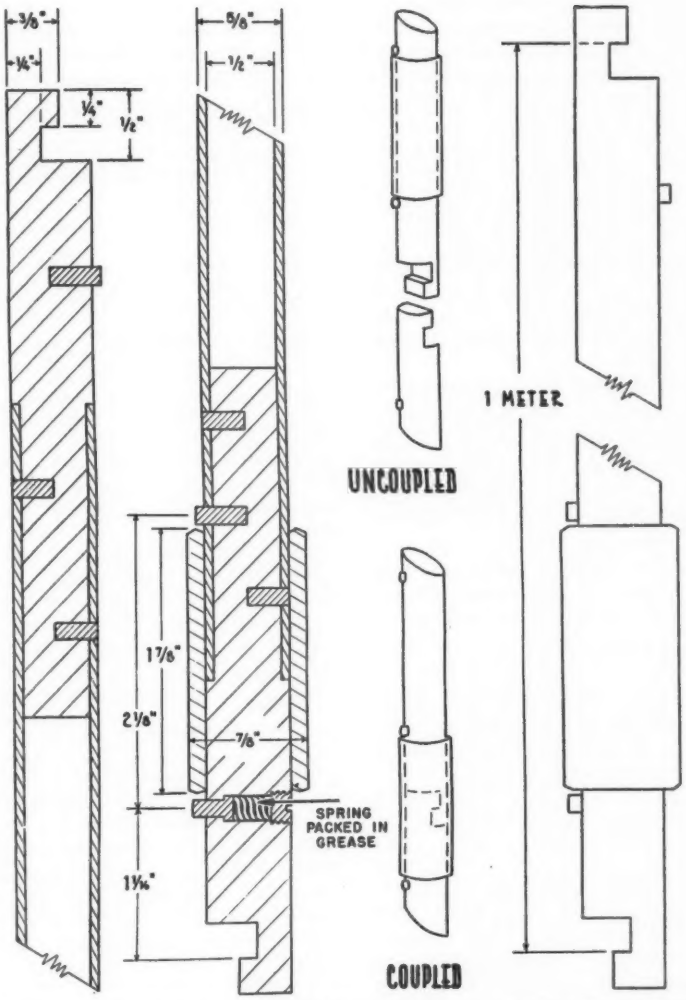

FIGURE 3.—IMPROVED LIGHT-WEIGHT EXTENSION RODS FOR SWEDISH HILLER BORER

Finally, samples were taken to a depth of 8 meters in a peaty area about 3.75 kms west of Chapingo (Profile 10). Here the solidified material was encountered at a depth of 1.50–1.75 m, the layers above it having been somewhat shrunken by drainage and cultivation, with resultant damage to pollen near the surface.

All the material collected in 1949 was brought back to Oberlin and analyzed.

1950

Borings were made at Culhuacán (Nahua) (Profile 3) and Pueblo San Lorenzo not far from Tlatilco, an Archaic site. About 5 m of samples below the base of the Nahua pyramids of Santiago Tlatelolco was also obtained from Doctor Zeevaert (Profile 4).

The remainder of this season was spent in

reviewing various archeological sites, in conferences, and in securing and packing samples from a core taken by Doctor Zeevaert to a depth of over 70 meters under the Palacio de Bellas Artes.

Procedure

The samples obtained in 1948 I analyzed in Mexico in the laboratory of the Instituto de Microbiologia of the Instituto Politecnico which Professor Marroquen had placed at my disposal. Probably because this material was of Nahua age and fairly fresh, the simple KOH method sufficed.

The 1949 collection, analyzed at Oberlin, proved somewhat more refractory; samples were considerably older (Archaic) and hence probably more highly mineralized. Pretreatment with HCl, cold to dissolve calcium salts and hot to remove those of magnesium, was found necessary, after which standard KOH treatment to loosen organic floccules gave good results. Even more drastic treatment has been found necessary for the deeper sediments, as will be reported in later papers.

A minimum of 100 pollen grains from each level was counted, except where the symbol @ followed by the number found indicates a lesser count. Frequent checks were run by counting additional hundreds.

Results

Tables 1–10 show percentages of the four forest genera, *Abies* (fir), *Pinus* (pine), *Alnus* (alder), and *Quercus* (oak). Their ecological relationships and probable indicator value have been discussed and shown in the diagram (Fig. 2). After some experience, supported by elementary statistical treatment, three types of sample were set up, as follows:

(1) Those in which the pine pollen constituted 90 per cent or more. These were assumed to indicate conditions too dry to sustain much oak, alder, or fir, all of which are relatively mesophytic as compared to the pines, where the latter are taken as a group.

(2) Those in which the oak pollen equalled 10 per cent or more. These were assumed to indicate relatively moist conditions.

Table 1.—Pollen Profile, Lake Chalco

About 200 yards west of school at Xico near ditch junction. Former chinampa, evidence of mixing to 160 cms. Nahua occupation, high oak, moist climate.

Depths cms	*Abies*	*Pinus*	*Alnus*	*Quercus*	
20	02	48	04	46	@82
60	00	82	03	15	@66
120	02	86	01	11	
160	02	78	10	10	
200	05	74	05	16	
210	03	71	11	14	

Table 2.—Pollen Profile, About 5 Meters North of Profile in Table 1

Former chinampa. Nahua occupation, high oak, moist climate in upper two samples. Transition at 200 cms. Pollen absent at 220 and 240 cms, probably due to oxidation during dry-climate low-water level, and at 260 cms volcanic ash probably correlative with that overlying Late Archaic in subsequent profiles.

Depth cms	*Abies*	*Pinus*	*Alnus*	*Quercus*	
80	02	74	05	19	
140	04	65	12	19	
200	06	87	03	03	@33
220 nil					
240 nil					
260 nil volcanic ash					

Table 3.—Pollen Profile, Culhuacán, Beginning just Below Zone of Nahua Artifacts

Note high oak, humid climate preceding these signs of occupation, zone of scanty pollen at 260, indicating dryness, ash, or both. High pine dry climate from there to bottom, where high oak, humid climate reappears. Analysis by Kathryn Clisby.

Depth cms	*Abies*	*Pinus*	*Alnus*	*Quercus*
120	00	70	15	15
175	00	66	17	20
222	00	74	05	21
260		Tr	(Ash?)	
290	00	97	03	00
320	00	98	00	02
350	01	91	02	06
380	02	89	06	03
395	02	90	01	07
410	02	91	02	05
440	00	88	06	06
470	03	74	10	13

TABLE 4.—POLLEN PROFILE, SANTIAGO TLATELOLCO, BEGINNING JUST ABOVE BASE OF OLDEST PYRAMID

Showing typical high oak, moist Nahua climate Collected and analyzed 1948. Remaining samples from core by Zeevaert, below pyramid, 1950. Note dry interval between bottom and top samples, also ash above pottery at 770. Pottery should be Late Archaic—if so, one of rare instances where it is overlain by Nahua. Charcoal probably inwash following volcanic fires.

Depth cms	*Abies*	*Pinus*	*Alnus*	*Quercus*	
460	12	50	08	30	@24
550–582 (gritty, no pollen)					
582–640 (charcoal and muck, no pollen)					
662–710 (gritty, volcanic shards, no pollen, evidently ash)					
770 (fragments of pottery and bone, no pollen)					
834	01	94	01	04	
888	00	93	06	01	
916	02	89	01	08	
978	01	82	04	13	

TABLE 5.—POLLEN PROFILE, EL TEPALCATE, RUSH TURF NEAR EDGE OF WATER

Note position of coarse sandy ash which, a few meters south, overlies about 55 cms of Late Archaic (Teotihuacán I) artifacts. Late Archaic occupation, high pine, dry climate.

Depth cms	*Abies*	*Pinus*	*Alnus*	*Quercus*	
25 (Black muck)					
50 (Black sandy muck)					
75 (Coarse volcanic ash)					
100 (Correlative of Late Archaic Zone)					
125	00	96	02	02	@93
150	00	100	00	00	
175	00	99	01	00	
200	02	97	01	00	
225	01	96	00	02	
250	01	97	00	02	

(3) All others—*i.e.* those with pine less than 90 per cent and oak less than 10 per cent. These have been treated as indecisive. Actually we have found that alder increases with oak as a rule. If we had lumped these two genera against pine, our tables would show even more striking consistency than they do. Fir, which rarely exceeds 5 per cent of the total, comes from elevations above 2800 meters and has given us some trouble in recovery, so that it has not

TABLE 6.—POLLEN PROFILE, CHIMALHUACÁN, 1 KM. SOUTH OF EL TEPALCATE

No signs of erosion or disturbance. Lake until about 1930. High oak, moist climate at top. Zone of exposure and oxidation at 50 cms. High pine, dry climate 75 cms and below. Coarse solidified volcanic ash encountered at 220 cms.

Depth cms	*Abies*	*Pinus*	*Alnus*	*Quercus*	
0	00	41	10	49	@39
25	00	47	06	47	@35
50		tr		tr	
75	01	97	02	00	
100	00	100	00	00	
125	01	99	00	00	
150	02	98	00	00	
175	02	98	00	00	
200	01	99	00	00	
210	04	96	00	00	
220	02	98	00	00	@50
220–240 Coarse solidified volcanic ash					
240 Correlative with Late Archaic Zone					

TABLE 7.—POLLEN PROFILE ABOUT 90 METERS NORTHEAST OF PROFILE IN TABLE 6

See Table 6 for explanation.

Depth cms	*Abies*	*Pinus*	*Alnus*	*Quercus*	
12.5	01	49	00	50	@14
25 nil					
37.5 nil (Temporary exposure and oxidation)					
50 nil					
62.5	05	95	00	00	@40
75	01	98	00	01	
87.5	00	100	00	00	
100	06	92	01	01	
112.5	03	96	00	01	
125	03	97	00	00	
137.5	02	98	00	00	
150	02	98	00	00	
167.5	02	98	00	00	
175	03	96	00	01	
187.5	03	97	00	00	
200	04	95	00	01	
212.5	01	99	Coarse solidified volcanic ash		
220	02	98	Coarse solidified volcanic ash		
(Correlative with Late Archaic.)					

TABLE 8.—POLLEN PROFILE NEAR SITE IN TABLE 6, BUT CONSIDERABLY DEEPER

High pine, dry climate below ash shows only at 300-cm level. From 325 through 675 cms an almost continuously high oak, moist climate, to be associated with the dark soil and high lake levels of Early Archaic. From 725 through 825 cms a still earlier high pine, dry climate.

Depth cms	*Abies*	*Pinus*	*Alnus*	*Quercus*	
25	00	39	11	50	@36
50 (Zone of temporary exposure and oxidation)					
75	00	100	00	00	@32
100	01	99	00	00	
125	00	100	00	00	@16
150	04	96	00	00	
175	03	97	00	00	
200	01	99	00	00	
225 (Coarse solidified volcanic ash)					
245 (Correlative with Late Archaic Zone)					
300	01	90	03	06	
325	02	87	00	11	
350	01	88	01	10	
375	00	83	01	16	@77
400	01	74	04	21	
425	01	85	04	10	
450	04	90	02	04	@80
475	01	89	03	07	
500	00	81	05	14	
525	00	85	02	13	
550	00	77	08	15	
575	00	84	01	15	
600	00	81	04	15	
625	00	82	09	09	
650	00	87	04	09	
675	00	80	06	14	
700	00	94	02	04	@93
725	00	94	01	05	
750	00	98	00	02	
775	00	95	03	02	
800	00	96	02	02	
825	00	90	01	09	

been used as a basis of interpretation at this juncture.

For convenience, the percentages we do consider decisive have been underscored, and general results summarized in Figure 4. The tables show that the two indicator types of spectrum fall into consistent sequences. On the other hand, the profiles obtained from the mounds at Xico, where the strata had been disturbed and mixed, are much more erratic. The Xico analyses here published (Profile 1 and 2) are from cores in the open lake bed, away from the mounds.

TABLE 9.—POLLEN PROFILE, LAKE TEXCOCO, DRY BED 3.75 KM SOUTHWEST OF CHAPINGO

Ash layer that covers Late Archaic at El Tepalcate has been exposed by wind deflation. Same sequence as in Table 8, but more condensed.

Death cms	*Abies*	*Pinus*	*Alnus*	*Quercus*
5 20 30 (Coarse solidified volcanic ash) (Correlative with Late Archaic)				
50	00	98	02	00
75	02	86	04	08
100	00	93	04	03
125	00	76	06	18
150	00	85	01	14
175	00	89	04	07
200	00	86	02	14
225	00	86	00	14
250	00	79	04	17
275	02	94	00	04
300	00	91	03	06

TABLE 10.—POLLEN PROFILE, CHAPINGO, PEATY CULTIVATED FIELD 3.75 KM WEST

Note agreement with Table 8, several kilometers distant, though evidence of dry period at 600 cms and below is less striking. This may be due to faulty technique, since our handling of older sediments was not perfected until after these analyses were made.

Depth cms	*Abies*	*Pinus*	*Alnus*	*Quercus*	
20 40	00	36	21	43	@28
75 (Zone of temporary exposure and oxidation)					
125	01	98	01	00	
140	01	89	05	05	
150 (Coarse solidified volcanic ash) 175 (Correlative with Late Archaic Zone)					
225	00	99	00	01	
275	00	87	03	10	
325	02	93	00	05	
375	01	83	01	15	
425	00	85	04	11	
450	01	88	01	10	
475	00	85	03	12	
500	00	75	07	18	
525	00	72	04	24	
550	01	83	04	12	
575	00	85	03	12	
600	01	94	00	05	
625	00	88	03	09	
650	00	89	03	08	
675	01	84	07	08	
700	00	85	03	12	
725	01	94	00	05	
750	01	87	03	09	
775	00	91	00	09	
800	00	83	03	14	

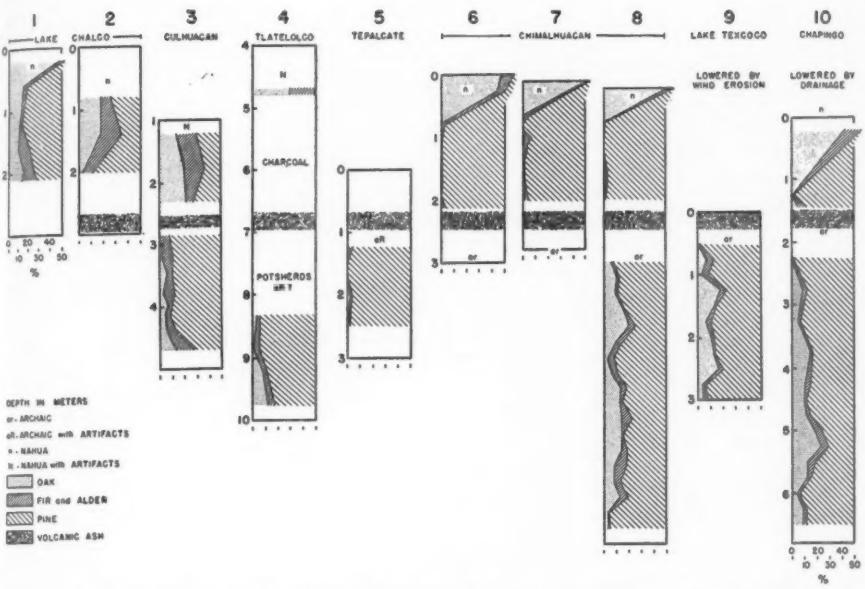

FIGURE 4.—TEN PROFILES SHOWING RELATION BETWEEN POLLEN FREQUENCIES OF CLIMATIC INDICATORS AND KNOWN ARCHEOLOGICAL LEVELS

Little or no pollen is preserved in layers, such as soils and beaches, which have been exposed to the air. Pollen is also scarce or lacking in volcanic ash deposits. Human occupation levels, in addition to the artifacts they contain, show under the microscope an abundance of charcoal. By including the main sandy layer, which I consider to be an ash fall, and the two identified cultural zones in the pollen columns of Figure 4, a simple and straightforward correlation is secured, as follows:

Nahua occupation— high oak—moist climate A
(Later Teotihuacan)
(Early Teotihuacan)— high pine—dry interval B_1
sandy layer (volcanic) B_2
Late Archaic-Teotihuacan I—
high pine—dry interval B_3
(Early and Middle Archaic)—
high oak—moist interval C
high pine—dry interval D

This sequence is further confirmed by the upper 10 meters of a 68-meter core which will be described in a subsequent report.

DISCUSSION

Two archeological zones, Nahua and Late Archaic, are unbracketed because their stratigraphic position has been directly established by artifacts *in situ*.

The Early and Middle Archaic, however, may be assigned to moist interval C, because:

(1) The classical Middle Archaic site of Zacatenco occurs on a beach at 2247 meters elevation.

(2) All known Archaic sites except the Late occur at elevations above the present basin floor.

(3) Archaic culture at Cuicuilco (lava-covered) and at Copilco occupied dark organic soils.

Evidently then the Archaic culture began at high elevation during a period of favorable moist climate and persisted into the subsequent dry interval B_3 during which it descended to the low level of El Tepalcate until obliged to move perhaps by the ash fall B_2.

Since the ash fall occurred during a dry period, the recent identification of Early Teotihuacán with Late Archaic indicates that the northward movement beyond the region of volcanic disturbance took place under dry conditions, a fact already suggested by the then low lake level. This removal was to ground nearly 30 kilometers north and more than 20

meters higher. Yet the higher and presumably drier site permitted not only survival but an impressive cultural development. Since our analyses show that dry conditions preceded and persisted for a considerable time after volcanic activity, how did Teotihuacán obtain its necessary water supply?

The most reasonable surmise is that sufficient reserves of ground water, probably including springs, had been safe-guarded by forests then covering the adjacent hills. Such forests would insure that any surface drainage would be usable and controllable. One may still see on the road from Mexico City to Teotihuacán, a clear, vigorous stream fed from ground water. Very probably such sources once existed at higher levels and in greater abundance than now.

A somewhat parallel instance can be found at the dried lake bed known as the San Augustin Plains in New Mexico. Bat Cave was occupied 4000 years ago, but artifacts around the upland springs are all fairly recent. A lacustrine culture has no choice but to move upward when the lake dries out.

Teotihuacán collapsed as a cultural center about 900 A.D., doubtless after the dry period had ended and after the beginning of the humid climate which favored later Nahua development. Vaillant suggested that this collapse was connected with deforestation for the burning of lime, and Cook reported evidence of heavy erosion in the Teotihuacán area. Apparently not even more favorable moisture conditions could relieve the effects of exploitation of the land. Certainly this would agree with experience elsewhere. We know that late Teotihuacán structures show evidence of elaborate means for intercepting rainfall (Linné, 1942).

A surviving mural fragment (Caso, 1942) shows clearly that the Teotihuacán Paradise was a place of abundant water whose chief source was a fountain, the mouth of Tlaloc, Lord of Earth and Water. As the Eskimo Hell is cold and the Mediterranean Hell hot, so the general idea of a Paradise involves comforts and delights hard to come by in this world. Needs for domestic fuel can be met in a variety of ways—*e.g.*, by the use of grasses or desert shrubs. There is no substitute for water. If Vaillant's idea of the effect of deforestation is applied not to fuel but to water supply, the latter may well have been the limiting factor for this impressive culture, as it has been for other activities within the Basin of Mexico.

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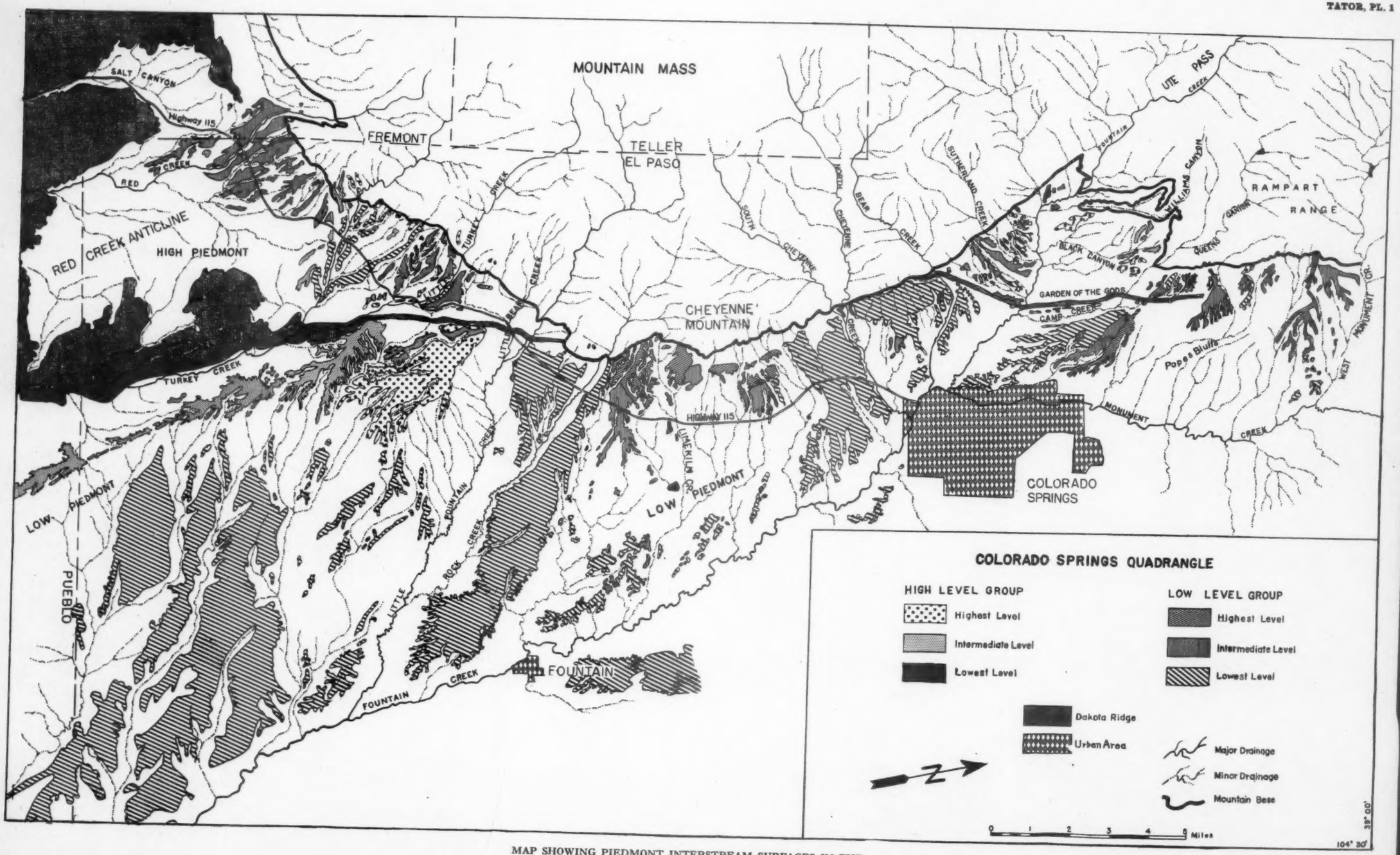

MAP SHOWING PIEDMONT INTERSTREAM SURFACES IN THE COLORADO SPRINGS QUADRANGLE AREA

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PIEDMONT INTERSTREAM SURFACES OF THE COLORADO SPRINGS REGION, COLORADO

BY BENJAMIN A. TATOR

ABSTRACT

The Colorado Piedmont, adjacent to the Front Ranges, contains broad interstream surfaces which stand above the modern valley bottoms, but in striking contrast to the steep mountain front and the aligned hogbacks of the foothills. The interstream flats occur both in abutting and detached relationship with the mountain front. Often a multiple arrangement is displayed, and several steplike levels are preserved. Merging of adjacent levels is common, particularly upstream. Remnants of higher surfaces, as well as ridges of resistant rock, project above the surfaces. On many, old stream channels converge toward a major canyon in the mountain front.

The piedmont surfaces are fanlike terraces floored with coarse fanglomerate deposits. The alluvium overlies bedrock erosion surfaces which truncate the tilted strata of the foothills belt, often extending onto the rocks of the mountain mass. Ordinarily the erosion surfaces disregard variations in rock type. The chief irregularities in the bedrock surfaces are broad, shallow depressions and small, channel-like incisions.

Wide, thinly alluviated valleys with relatively flat bedrock floors occur below the piedmont interstream surfaces. The width of these valleys correlates with the resistance of the rock types traversed. Some of the valleys originate in the mountain mass and extend through the piedmont, whereas others, tributary to the mountain-born drainage lines, begin in the piedmont. Most are dry throughout the year, except during occasional summer storms. Valley segments in the mountain area are deep and narrow, except locally where small, thinly alluviated, flat-floored basins are present (Tator, 1949, p. 1774–1783[1]). The piedmont valleys contain flights of rock-floored terraces which form steps below the interstream surface levels.

The origin of the piedmont levels is a major key to the diastrophic history of the Rocky Mountain front region. This present work includes a brief summation of concepts of origin of similar land forms in other areas, detailed descriptions of piedmont interstream surfaces in a selected portion of the piedmont, a brief analysis of the processes of valley widening by sidewall weathering and rock planation by vertical corrasion now active in the modern valleys (Tator, 1949), an adaption of these processes to development of terrace and piedmont interstream surfaces in the region, an interpretation of the physiographic history of the area studied, and tentative establishment of middle or late middle Pleistocene and Recent as the intervals during which development of the various surfaces took place.

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[1] This work describes processes responsible for origin of these basins.

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Introduction

Location

Field research for this work was confined to the Colorado Springs quadrangle along the Front Range in east-central Colorado, including portions of El Paso, Teller, and Pueblo counties (Fig. 1). The piedmont surfaces here are well developed, unusually well preserved, and their pattern of distribution shows clear relationship to the modern valley bottoms.

Field Methods

The base map was compiled from U. S. Forest Service aerial photographs flown in 1937. Recent topographic sheets of the Army Map Service were also valuable in this respect; including, of Series V877, sheets 5061, III, SW (Mount Pittsburg); 5061, III, SE (Timber Mountain); 5061, III, NE (Cheyenne Mountain); 5061, III NW (Mount Big Chief); and 5061, IV, SE (Colorado Springs). Plate 4 shows enlargements of portions of these maps. Elevations and sections were measured with Paulin Altimeter, and certain key areas were mapped with plane table and alidade. This is the second of a series of papers resulting from geomorphic studies carried on during the summers of 1940–1941 and 1946 to 1951, and the winter of 1945–1946.

Previous Work

The geomorphology of the Colorado Springs quadrangle has not heretofore been studied in detail. Hayden (1874) noted broad alluvial deposits, derived from the mountain rocks, on the eastern mountain flank near Colorado Springs. Finlay (1916) concluded that the gravel deposits once formed a continuous surface. Van Tuyl and Lovering (1935) indicated erosion surfaces on benches and spurs in the Cheyenne Mountain vicinity south of Colorado Springs, concluding that Pliocene pedimentation accounted for the higher levels. The present paper describes in detail the surfaces in question.

Acknowledgments

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Concepts of Origin of Bedrock Erosion Surfaces

The origin of bedrock surfaces similar to those described here has been controversial since Gilbert (1877) first recognized them. Various terms have been used for such surfaces, including pediment, conoplain, suballuvial bench, rock fan, rock plane, and peripediment. Two major theories of origin have been evolved: one attributes development to the planation effect of laterally migrating streams; the other emphasizes escarpment retreat under the influence of the various processes of subaerial weathering.

McGee (1897) used the term "pediment" for the rock plains of southwestern Arizona and Mexico, relating origin to sheet-flood erosion. Johnson (1932), Field (1935), Mackin (1948), and Howard (1942). Johnson, a strong exponent of lateral corrasion, maintained that

FIGURE 1.—DRAINAGE MAP OF COLORADO SPRINGS REGION

Davis (1905) at first preferred McGee's postulate, though he later accepted other possibilities. Tolman (1909) showed that the sheetflood is a depositing rather than an eroding agent, more a result of a planate bedrock surface than a cause of its development.

Gilbert (1877) explained the surfaces of bedrock erosion in the Henry Mountains, Utah, as the product of lateral corrasion by graded streams. Many have agreed with this idea, including Lee (1900), Ogilvie (1905), Paige (1912), Bryan (1922), Blackwelder (1931), mountain-front retreat results largely from undercutting by laterally swinging streams that leave a plane of truncation at the mountain base. He used the terms "rock fan" and "rock plane" to designate the surface thus produced.

Other workers believe that the bedrock planes are developed along the bases of retreating escarpments, which are caused mainly by weathering. Lawson (1915) contended that the piedmont surface, "the suballuvial bench," is developed as a surface of transportation between the retreating mountain escarpment and

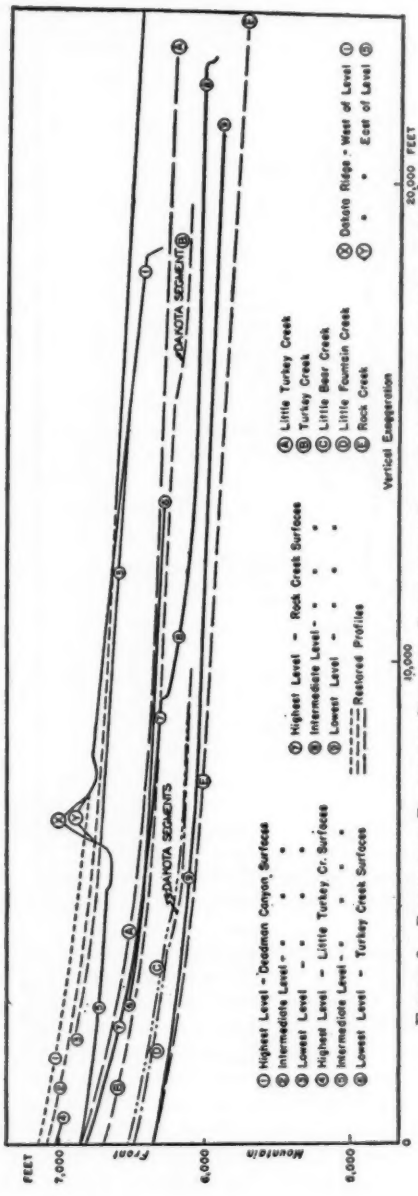


FIGURE 2.—PROFILES OF DIAGNOSTIC PIEDMONT INTERSTREAM SURFACES AND PIEDMONT STREAM SEGMENTS

filling basin. Davis (1930) essentially agreed with this concept. Mortensen (1927), Waibel (1928), Bryan (1940), Glock (1932), and Rich (1935) held somewhat similar views.

The more recent tendency has been to accept a combination hypothesis. Bryan (1925) recognized both lateral planation and weathering of escarpments in the production of extensive rock planes. Gilluly (1937) believes that lateral corrasion, rill wash, and weathering with removal of the weathered debris by various agencies, all operate conjunctively, though any one may dominate at a particular locality. Sharpe (1940) found that some portions of the bedrock surfaces developed under lateral planation, whereas other portions were produced by weathering and slope transportation.

General Setting

Geology

Sedimentary strata underlie the piedmont, and older igneous and metamorphic rocks form the mountain slopes. The stratigraphic section extends from Precambrian through Upper Cretaceous. A thick sedimentary section dips eastward on the mountain flank forming a homocline. Folding in the southwest portion has produced a broad, southward-plunging anticline (Glockzin and Roy, 1945). In some localities thrust faulting has steepened dips of strata adjacent to the mountain base, causing overturning and cutting out segments of the normal stratigraphic sequence (Pl. 1). A structural embayment of sedimentary rock, the Manitou Embayment, occupies a westward indentation of the mountain front west of Colorado Springs.

Topography

The mountain front presents an abrupt wall-like feature bordering the piedmont, with the steepest portion along the trace of the Cheyenne Mountain thrust fault near the center of the quadrangle (Pl. 1). Here resistant igneous rocks (Precambrian) overlie weak sedimentary strata (Upper Cretaceous). The more resistant beds of the foothills belt include the Fountain (Pennsylvanian), Lyons (Permian), and Lakota-Dakota (Cretaceous) formations. Each forms

a hogback except where eliminated by faulting. The most prominent ridge, formed by the Lakota-Dakota unit, creates a distinct physiographic boundary between high and low sections of the piedmont. Higher portions are contained between this ridge and the mountain escarpment, lower elevations occur east and southeast of the ridge. Three areal units are definable: the *Mountain Mass*, underlain by resistant igneous and metamorphic rocks; the *High Piedmont*, on relatively resistant Paleozoic and Mesozoic strata; and the *Low Piedmont* whose surface is formed on less resistant Late Mesozoic rocks, for the most part Pierre shale (Upper Cretaceous) (Fig. 1).

Drainage

The drainage lines within the Mountain Mass occupy deep, narrow canyons with bedrock floors, widened in places by thinly alluviated basins developed by valley-widening processes (Tator, 1949). The mountain valleys have segmented profiles, steepen noticeably through some stretches, and shallow in others. The most abrupt gradient changes occur at the mountain base. Red Creek, in the southwest portion of the quadrangle (Fig. 1), shallows very abruptly from 160 feet per mile in its canyon mouth to 100 feet per mile in the High Piedmont below. Similarly, Turkey Creek slopes at 125 feet per mile above its canyon mouth, changing to 75 feet per mile in the High Piedmont. A 200-foot per mile gradient in the canyon segment of Little Fountain Creek becomes 150 feet per mile in the piedmont (Fig. 2).

The elevations of the canyon mouths of the streams also differ markedly. Little Turkey Creek, for example, enters its piedmont segment at 6825 feet, whereas Little Bear Creek, approximately a mile to the north, is at 6550 feet at a similar point in its profile (Fig. 2). Furthermore, canyon-mouth elevations increase along the mountain front both northward and southwestward from Rock Creek. Figure 2 shows the southwestward increase; the northward increase is revealed by comparing the mouth of Rock Creek canyon at 6350 feet with West Monument Creek canyon (north boundary of the quadrangle) at 7200 feet (Pl. 1).

Three major mountain-born drainage lines—Red, Turkey, and Little Fountain creeks—escape from the High Piedmont via water gaps in the Dakota ridge (Fig. 1). Each steepens rapidly in gradient through its short watergap segment (Fig. 2). Thus, Red Creek, with a slope of 25 feet per mile in the High Piedmont, attains 50 feet per mile in crossing the Dakota unit; Turkey Creek, at 100 feet per mile in the High Piedmont, rapidly steepens to 250 feet per mile through the ridge, shallowing to 75 feet per mile in the Low Piedmont; Little Fountain Creek has a gradient of 150 feet per mile upstream from the ridge, increasing to 200 feet per mile traversing the Dakota and shallowing to 100 feet per mile just below in the Low Piedmont (Fig. 2).

Faulting cuts out the ridge-forming Dakota unit a short distance north of Little Fountain Creek, but it reappears in the embayed area west of Colorado Springs (Pl. 1). In the intervening stretch of country below the mountain base, stream segments maintain relatively smooth profiles through the weak rocks (Pierre shale) of the Low Piedmont. Rock Creek (Fig. 2) best exemplifies this condition. Fountain Creek, master stream of the quadrangle, crosses the Dakota unit upon leaving the embayment west of Colorado Springs, exhibiting gradient characteristics similar to those of Dakota-controlled streams farther south. The gradient increases from 40 feet per mile upstream to 100 feet per mile through the ridge segment, then decreases to 50 feet per mile in the Low Piedmont.

Climate

This semiarid region displays rather specialized types of geologic activity along the modern drainage lines. The mean annual rainfall ranges from 12 to 14 inches; more than 75 per cent falls between April and September. Heavy localized precipitation and torrential runoff are characteristic. The active geologic processes of the valley bottoms are controlled by the existing climatic regime.

Piedmont Interstream Surfaces

The piedmont interstream surfaces can be separated into High-Level and Low-Level groups, each including several related levels. Levels within each group tend to merge moun-

tainwards, with a less pronounced merging in the same direction among the groups. In general, mountain-base elevations of surfaces in each category increase northward in the quadrangle. Local differences in elevation of comparable levels are largely due to the Dakota ridge, although other causes prevail, such as: (1) extent of mountainward development, levels projecting farthest into the Mountain Mass being the highest; and (2) differences in elevation of the mountain canyon segments of the stream valleys where the surfaces formed. Differences in elevation of modern canyon-mouth segments have been pointed out (Fig. 2). Outstanding examples of elevation differences characterize Low-Level surfaces in the High and Low Piedmont areas, particularly in the southern half of the quadrangle (Fig. 2).

High-Level surfaces, except in certain localities, are of limited extent, but remnants are scattered throughout the three areal units. Diagnostically, these have the greatest mountain-base elevations, from 6950 to 7500 feet, and the thickest alluvial sections. The largest remnant, the Deadman Canyon surfaces, is located on the drainage divide between Little Fountain and Turkey creeks. Less extensive representatives of the group occur within and adjacent to the mouths of major canyons in the mountain front, usually as flat-topped elements on mountain spurs. The best-preserved remnants in the northern portion of the quadrangle are flat-crested ridges between Black and Williams and Black and Queens canyons at the south end of the Rampart Range (Pl. 1).

Low-Level surfaces are present throughout the quadrangle (Pl. 1) and remnants are more widely distributed than those of the High Level. Representatives occur in the High Piedmont at mountain-base elevations of 6650 to 7050 feet; in the Low Piedmont, apex elevations range from 6300 feet near Rock Creek to 7300 feet at the northern boundary of the quadrangle. The most extensive Low Piedmont remnants are preserved along the north side of Rock Creek and on the interfluve between Camp and Monument creeks northwest of Colorado Springs (Pl. 1). Prominent levels in this group are found along Turkey and Red creeks in the High Piedmont.

The stream terraces, present within the vertical interval between piedmont interstream surfaces and the modern valley bottoms, include several levels. These merge with the valley bottoms upstream in the vicinity of the mountain base. The highest terrace is usually closely related to the adjacent interstream flat and the two tend to merge upslope. The piedmont surfaces and the terraces are similar in every respect, save that the terraces are much more restricted in extent.

Deadman Canyon Surfaces

General Description

These surfaces form a flat interfluve between the Little Fountain and Turkey creek drainages, extending for 12 miles along the eastern edge of the High Piedmont (Pl. 1). The alluvium in southward continuations of the surfaces in the Pueblo quadrangle was described by Gilbert (1897) as the Nussbaum formation. The Dakota ridge flanks the western edge of the northern portion for 2 miles (Pls. 1, 4); the continuity of the ridge is interrupted by the Turkey Creek water gap for some distance to the south. Three levels are contained in the remnant, two along the east side of the Dakota ridge and a third in the southeastward continuation along Turkey Creek. The highest level forms most of the surface of the northern

PLATE 2.—ALLUVIAL SECTIONS IN ROCK CREEK AND MESA SURFACES

Figure 1.—Alluvial-Filled Channel in the Bedrock of the Middle Level of the Rock Creek Surfaces

View is east into the road cut of Highway 115. The channel extends about 15 feet below the average bedrock plane. The bedrock is gently eastward-dipping Pierre shale. The coarse texture of the fanglomerate is apparent.

Figure 2.—Alluvial Section in the Mesa Surfaces

Approximately 12 feet of a 40-foot section in the Colorado Springs city gravel pit is visible. The fairly well bedded, well sorted zones are quite typical of alluvial deposits in the distal portions of the piedmont interstream surfaces.

FIGURE 1

FIGURE 2

ALLUVIAL SECTIONS IN ROCK CREEK AND MESA SURFACES

FIGURE 1

FIGURE 2

LOW-LEVEL SURFACES IN THE HIGH PIEDMONT AREAS

portion (Pl. 4), is fan-shaped, and slopes from 6760 feet toward the southeast at 100 feet per mile. This level is also preserved to the north along the mountain front near Little Bear and Little Fountain creek canyons, on flat-topped spurs at 7150 feet. The gradient of restored profile extending to the mountain front is 250 feet per mile (Fig. 2).

A second level forms the surface of the southeastward continuation along Turkey Creek, sloping from 6630 feet at the south end of the Dakota ridge with a gradient of 100 feet per mile (Pl 1.). Ridges and butte like remnants preserve this surface mountainward along Turkey and Little Turkey creeks. It also occurs in fragmentary form on mountain spurs around the canyon mouths of these streams at 7100 feet. The restored profile mountainward has an average gradient of 250 feet per mile (Fig. 2).

The flat aspect of each of these levels is diversified by broad shallow swales which form a radiating system representing old stream channels. On the northern portion of the surface, these converge toward Little Fountain Creek canyon. The largest and deepest parallels the western base of the Dakota ridge, gradually widening downslope for 2½ miles. The north-end elevation is 7630 feet, and the slope is southeast at 100 feet per mile (Fig. 2; Pl. 1). No remnants of this level have been identified mountainward from the Deadman Canyon surfaces. It is preserved in limited extent, however, along the eastern side of Turkey Creek valley as terrace-like features 40 to 50 feet below the higher level already described.

Alluvium

Alluvial thicknesses vary under the three levels of the Deadman Canyon surfaces. The northeast edge of the highest level exposes an alluvial section 90 feet thick at the west and 110 feet thick at the eastern ends. This section thins gradually through 3½ miles to the southeast where the section is 50 feet thick.

The next lower level is underlain by 40 feet of similar material just south of the Dakota ridge; it thins to 10 feet through the 10-mile continuation to the south. The old channel bottom in the northern portion of the surface has a 25-feet section in the north end and a 40-foot section in the southern extremity. No marked differences in cross-section thickness are apparent in the alluvium of any of these levels.

The fanglomerate is massive, though locally bedding and sorting are evident. It consists largely of coarse sand with cobbles and boulders up to 3 feet in diameter occurring in lenses and scattered throughout the deposit. Average grade sizes decrease, and the factors of rounding and sorting are augmented downslope. Thin, irregular lenses of pebble and cobble "mortar-beds" are common, particularly near the base of the material. Igneous and metamorphic rock fragments from the Mountain Mass constitute the major percentages, although minor amounts of sedimentary particles are intermixed. Of the latter type, Dakota sandstone is in the majority; 1- to 2-foot boulders are common.

The absence or presence of certain constituents aids in distinguishing the remnants of the three levels, and also in determining the position of the old drainage lines. The fanglomerate of the northern portion of the Deadman Canyon surface contains abundant rhyolite cobbles and boulders. Boss-like bodies of this rock are located in the Mountain Mass near Little Bear Creek, the only other occurrences are several miles northeast in Rock Creek canyon. The Little Bear Creek rhyolite ex-

PLATE 3.—LOW-LEVEL SURFACES IN THE HIGH PIEDMONT AREAS

FIGURE 1.—PIEDMONT INTERSTREAM SURFACE NORTH OF QUEENS CANYON

This remnant is confined between the mountain base (right of view) and the Lyons and Dakota ridges (left of view). The vista is south along the east side of the Rampart Range. The steeply tilte strata in the foreground are Fountain (Pennsylvanian) and, except for minor projections above the surface, are truncated by the erosion plane of the piedmont interstream surface.

FIGURE 2.—PIEDMONT INTERSTREAM SURFACE JUST NORTHEAST OF TURKEY CREEK

This is an erosion plane developed on gently eastward-dipping Fountain strata. Turkey Creek extends from right to left in the middle distance. The tree-covered ridge in the background is a remnant of the High-Level group along the southwest side of this stream. The low surface is a bedrock flat from which the alluvium has been removed.

posures are in the mountain area from which the old surface channels radiate. On the other hand, the absence of rhyolite fragments in remnants of the level along Turkey Creek valley supports a conclusion that a different drainage, without access to the rhyolite bodies, was the depositing agent. This view is bolstered by the general upslope trend of the Turkey Creek valley remnants toward Turkey Creek canyon.

Bedrock Surface

The alluvium rests on an erosion surface across the up-turned edges of eastward-dipping sandstones, limestones, and shales (Upper Cretaceous). This truncation surface is relatively planate; the chief departures from smoothness are broad shallow channels. The alluvial surface and the underlying bedrock plane slope in the same direction and at the same angle.

Comparison of Deadman Canyon Surfaces with Adjacent Surfaces

Surfaces of the Low-Level group are preserved at the edges of the Deadman Canyon surfaces in the High Piedmont to the west and in the Low Piedmont to the east and northeast (Pl. 1). Three lower levels, the Little Turkey Creek surfaces, occur as remnants along that stream west of the northern portion of the Deadman Canyon surfaces (Pl. 4). The highest, only a fraction of its former extent, is at 7000 feet and truncates steeply tilted sandstones close to the mountain base (Fountain-Pennsylvanian); the intermediate level, the most extensive of the remnants, follows the northeast side of Little Turkey Creek from 6870 feet at the mountain base and with a slope of 200 feet per mile; and the lowest level forms narrow terrace elements about 25 feet above the modern valley bottom of that stream. These levels are preserved throughout the High Piedmont (Pl. 1) in positions about 100 feet below remnants of the High-Level surfaces.

The northeastern edge of the Deadman Canyon remnant is a 600-foot escarpment leading down into the valley of Little Fountain Creek (Pl. 4). A broad Low-Level surface along the north side of this valley slopes northeast from 6350 feet at 150 feet per mile, and occupies the area between Little Fountain and Rock creeks east of the Dakota hogback (Pl. 1). This level is approximately 500 feet below the northern portion of the Deadman Canyon surfaces and 300 feet lower than the lowest Little Turkey Creek surface, with which it correlates. A slightly higher level is present on 20- to 30-foot knobs which project above the surface. The surfaces may be traced up the drainages of Little Fountain and Little Bear creeks as terraces.

Rock Creek Surfaces

General Description

These remnants of the Low-Level group flank the Mountain Mass in the Low Piedmont north of Rock Creek (Pl. 1). The uppermost is preserved in the surfaces of several fanglomerate-capped ridges; lower, southeastward continuations of the ridges form an intermediate level; and the lowest and most extensively preserved extends eastward from the mouth of Rock Creek canyon for 8 miles to the valley of Fountain Creek. The highest level is at 6650 feet at the mountain base just north of Rock Creek, rising to 6850 feet in a westward indentation of the foothills at Lime Kiln Creek, sloping eastward at 400 feet per mile (Fig. 2). Lower continuations eastward from 6350 feet below the scarped, distal edges of the highest level slope at 200 feet per mile (Fig. 2). The lowest level, at 6375 feet in the mouth of Rock Creek canyon, gradually decreases in slope from 200 to 50 feet per mile (Fig. 2).

Alluvium

Except for textural and lithologic differences, the alluvial material is similar to that in the Deadman Canyon surfaces. Exposures are excellent in Highway 115 road cuts which pass through the surfaces at locations half a mile to a mile from the mountain base (Pl. 4). Maximum thicknesses exposed range from 20 feet in the lowest level (Fig. 3) to 40 feet in the highest. The section thins both upslope and downslope from the road-cut exposures. Local minor variations in thickness occur.

FIGURE 3.—CROSS SECTIONS OF ROCK CREEK AND RED CREEK SURFACES
Along Colorado Highway 115. Location is indicated by corresponding lettering designations on Plate 4. Note the channel-like irregularities in the otherwise smooth bedrock surface. No vertical exaggeration.

Bedrock Surface

The bedrock surfaces are east- and southeast-sloping planes which truncate the homoclinal structure of the piedmont; the beveled strata are largely Pierre shale (Upper Cretaceous). Narrow channels extend 5 to 25 feet below the average bedrock level (Fig. 3; Pl. 2, fig. 1), and there are minor protuberances above this level. Despite these local irregularities, the bedrock plane is relatively smooth. The channels trend downslope.

Other Low-Level Surfaces

Mesa Surface

The most extensive remnant in the northern portion of the quadrangle, locally termed "The Mesa," extends for about 4 miles along the Camp Creek-Monument Creek interfluve (Pl. 1). Three levels constitute this surface: the highest, a narrow flat-topped ridge in the northeast portion; an intermediate, occupying the central portion; and the lowest, forming a broad flat on the west just above the valley of Camp Creek. These levels merge upslope toward the mouth of Queens Canyon; the slope in each case is south at from 200 feet per mile upslope to 100 feet per mile in the distal segments. The Mesa surface has, in addition, a moderate slope westward toward Camp Creek.

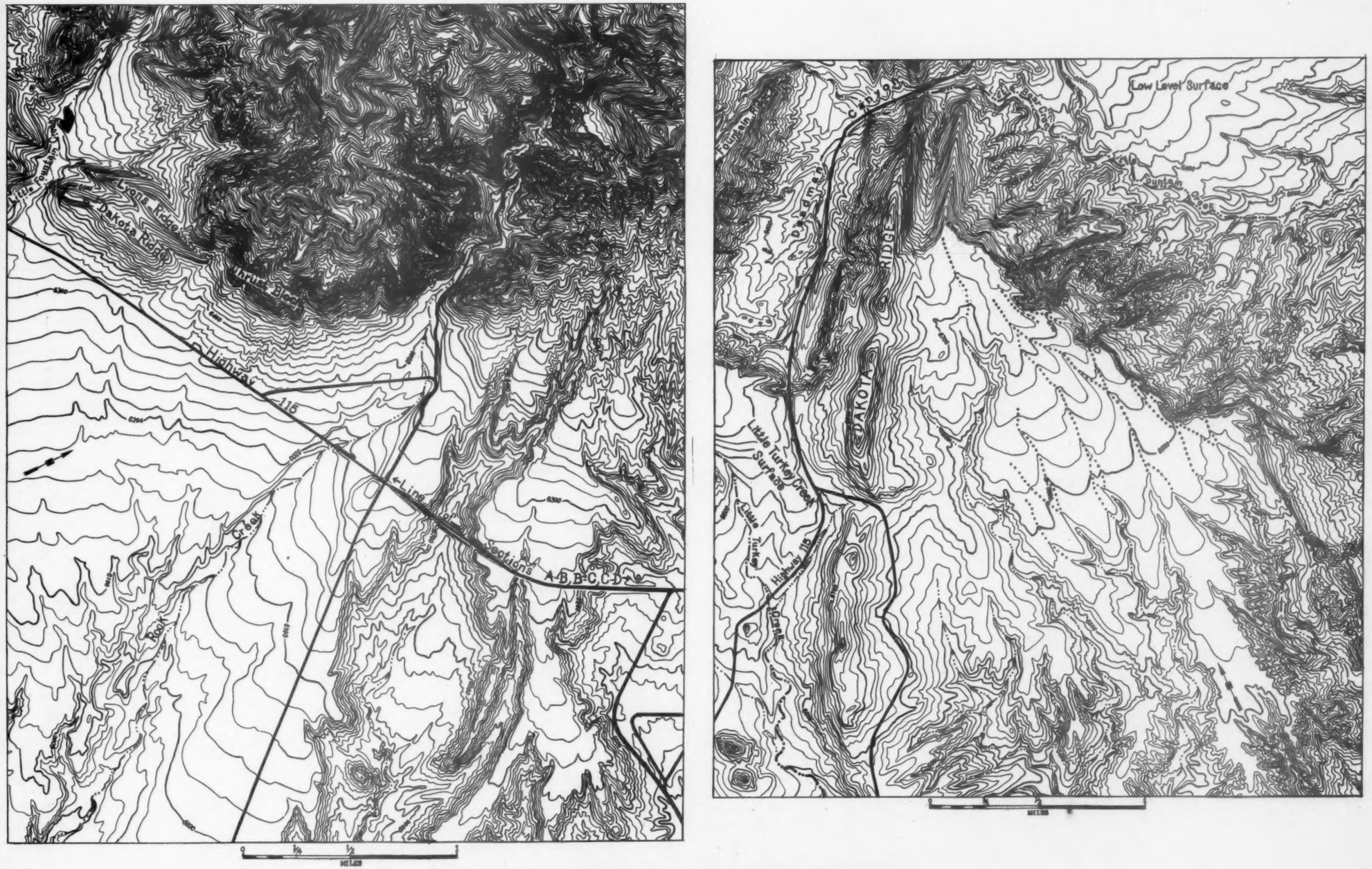
Broadly rounded swales radiate from the north end, representing old stream channels which once issued onto the surfaces from Queens Canyon. The alluvium is similar to that already described. A 35-foot section in the north end of the highest level thickens to 50 feet a short distance south but thins eventually to 10 feet under the southern portion of the remnant. Twenty-five feet of fanglomerate underlies the upper end of the intermediate level, thickening to 40 feet and then thinning again down the slope. The north end of the lowest level contains 40 feet; 3 miles south, the section thins to 10 feet. In direction transverse to the longitudinal slope, the alluvial section of each level thins toward the edges. The underlying bedrock surface, developed across the upturned edges of Pierre shale strata, is corrugated by broad, north-south trending sags which extend 15–25 feet below the average bedrock level. These sags widen and deepen down the slope, attaining several hundred feet in width in the distal portions. The Colorado Springs city gravel pit, exposing a 40-foot alluvial section, lies in the center of one of these sags (Pl. 2, fig. 2).

Lesser Surfaces

Multiple levels of the Low-Level group are present (1) along the mountain base north of the Mesa surface, (2) in flanking position at the base of Cheyenne Mountain, (3) between the Deadman Canyon surfaces and the valley of Fountain Creek in the southeast portion of the quadrangle, and (4) along Turkey and Red creeks in the High Piedmont (Pl. 1).

One particularly broad remnant extends east from a mountain base elevation of 6350 feet along the south side of Cheyenne Creek, and continues southeast for several miles as a terrace along Fountain Creek. Another, at 6450 feet at the mountain front, parallels the south side of Bear Creek. A number of east- and southeast-sloping levels flank the Rampart Range between the Mesa surface and the northern boundary of the quadrangle. One of these is confined between the Mountain Mass and the Dakota ridge a mile north of Queens Canyon. The slope, 500–600 feet per mile, is toward the Lyons and Dakota ridges, which abruptly terminate the surface to the east (Pl. 3, fig. 1). Three narrowly separated levels are contained, the highest with a mountain base elevation of 7100 feet. North from this locality, faulting has eliminated both of these ridge-forming units. Here a fan-shaped surface slopes eastward toward the base of Popes Bluff, a west-facing escarpment in resistant Eocene strata (Pl. 1). The greater longitudinal extent of this remnant can be explained by the absence of the restrictive influence of the Dakota ridge. The alluvium in this surface is much finer-textured than in those farther south; it consists of fragments of granite, similar to the pebble-size debris now being weathered from some portions of the Front Ranges.

Numerous short steep remnants occur south of Fountain Creek in the embayment west of Colorado Springs (Pl. 1). The slope in all cases is toward that stream in the same sense as the modern tributaries. The surfaces, composed



TOPOGRAPHIC MAPS OF ROCK CREEK AND DEADMAN CANYON SURFACES

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of several levels, are higher than their equivalents in the Low Piedmont to the east. The most extensive remnant parallels the east side of Sutherland Creek at 300 feet per mile.

The remnants along Turkey and Red creeks in the High Piedmont are similar to others described (Pl. 1). One, particularly well preserved, forms the western side of Red Creek valley from its canyon mouth southward for 3 miles. The slope is 200 feet per mile from a canyon-mouth elevation of 6700 feet. A less-pronounced slope is evident in transverse direction toward Red Creek. An alluvial section exposed along Highway 115 reveals 10 feet of fanglomerate under the western edge and 20 feet under the eastern edge (Fig. 3). The bedrock plane truncates steeply southeast-dipping Fountain strata (Pennsylvanian). The surfaces along Turkey Creek (Pl. 1 and Fig. 2) are less well preserved. Several are almost devoid of alluvium, but the exposed bedrock surfaces are remarkably smooth with only minor irregularities (Pl. 3, fig. 2).

Stream Terraces

Stream terraces in the valley sides at varying intervals below the piedmont interstream surfaces form flights of thinly alluviated, rock-floored levels. In the broader valleys, many terrace remnants are present, detached from the valley walls. The number of terraces is not uniform, in different valleys or in separate segments of the same valley. The maximum observed is seven; in some valleys none is present. In general, the number of terraces increases down-valley. Terraces in the mountain-canyon segments of the streams correlate down stream with piedmont interstream surfaces; none there are equivalent to modern terraces in the piedmont.

The vertical spacing between these levels is quite irregular. Clearly separable surfaces in one locality may merge elsewhere. They tend, however, to merge with the valley bottom in the up-stream direction, in sequence from lowest to highest. In addition, the vertical intervals between terraces tend to increase down-valley. On the average, in segments with distinct terracing, the vertical intervals are 15–20 feet.

The longitudinal slopes approximate those of the modern stream segments along which they occur. Many have a slight transverse slope toward the valley center, although a few slope toward the valley side. Paired terrace arrangements are rare in this region; most of them are along the larger stream valleys.

The alluvium in these surfaces is similar to the fanglomerate in the piedmont interstream surfaces. Thicknesses range from 5 to 20 feet. In general, this material forms a relatively thin, quite uniform blanket over a bedrock surface. The surface is identical to the bedrock plane of the piedmont surface, beveling the up-turned strata with little regard for rock differences. Several of the broader terrace remnants are indistinguishable from the piedmont interstream flat, save for being less extensive and closer to the modern drainage lines. Furthermore, all the features—abandoned channels, rock trains, and alluvial fans—of the modern valley bottoms are also present in the terrace surfaces (Tator, 1949, p. 1774–1776).

Geologic Processes

Slope Weathering

The rate of weathering on exposed surfaces of homogeneous rock is constant if the climate remains uniform and a means for removing the products of weathering exists. If processes of distintegration operate with equal persistence on all exposed rock surfaces, and the debris is removed, the surfaces will retreat uniformly. Furthermore, a constancy of slope angle will be maintained where similarity in climate, lithology, and rock structure prevails. These conclusions have been substantiated by Lawson (1915), Bryan (1940), Rich (1935), Field (1935), Gilluly (1937), and many others.

The Colorado Springs region contains diverse rock types and structures, and the effects on slope declivity and rate of slope retreat are often apparent. This is made particularly evident from a comparison of valley widths in the granite of the Mountain Mass, in the relatively resistant strata of the High Piedmont, and in the weak sediments of the Low Piedmont. Mountain segments are narrow, deep, and steep-walled, except where flat-floored, alluviated basins produce local widenings (Tator, 1949, p. 1774–1776). These basins in homo-

geneous rock are believed associated with local weaknesses produced by secondary rock structures, such as jointing, which form a locale for more rapid slope retreat than is normal to the mountain rocks. Valley segments in the High Piedmont are somewhat wider than those of the Mountain Mass, but narrower than those in the Low Piedmont. The rocks of the High Piedmont are less resistant than those of the Mountain Mass but more resistant than Low Piedmont strata. The most widespread piedmont interstream surfaces are developed in the weak shales (Upper Cretaceous) of the Low Piedmont, whereas the more restricted surfaces are in the more resistant strata, the sandstones and conglomerates of the High Piedmont (Pennsylvanian).

Channel Shifting

The mechanics of channel shifting in this region have been studied in detail (Tator, 1949, p. 1778–1781). The prevailing rainfall condition—intense localized precipitation which yields flood runoff of short duration but great intensity along limited valley segments—is largely responsible for channel-shifting activity. Aggradation by flood waters builds rock trains, boulder plugs, and main and tributary valley fans which, as obstacles, shift the runoff channels. Numerous abandoned channels in all parts of the valley attest to this activity. Impingement of channels against the base of the valley sides, forced into such positions by alluvial damming, is the means for removing debris accumulations. Thus, the side-walls are kept clean and permit uniform retreat. Locally, lateral corrasion may operate as a result of such impingement. As the sidewalls are weathered back uniformly, and the flood channels are shifted by alluviation, all portions of the valley bottom some time are the loci of channeling action.

Bedrock Erosion

The depth of effective scour by flood runoff is controlled by such factors as flood volume and velocity, the presence of channel-confining features on the valley floor, and the character of the material in the bottoms and walls of the channels. The floors of mountain-basin segments of valleys and piedmont valley segments in this region are thinly alluviated. In general, the valley-bottom deposit is uniformly spread over a relatively flat bedrock erosion surface. Variations in alluvial thickness are in the nature of fan deposits or filled channels which have incised the bedrock floor for short distances. The comparative flatness of the bedrock floor is the product of vertical corrasion to a controlled depth by the shifting flood channels. Precipitation is insufficient to promote continuous runoff, and the drainage is only intermittent. During runoff intervals, the torrents scour channels to bedrock; the channel shifting through long periods of uniform climatic conditions yields an all-pervasive effect. The fact that the bedrock floor is an erosion feature and relatively planate supports the contention that the scour depth remains a constant for long periods.

Relationship Between Valley Deepening and Valley Widening

The streams of the Colorado Springs region are actively degrading their valleys, as indicated by the deep, narrow canyon segments incised in the bedrock of the Mountain Mass, by the excessive steepness of the stream gradients, and by the bedrock exposures along the valley sides and beneath the broad flats of the piedmont area. In a degrading region such as this, the alluvium of the valley bottom is being continuously removed at a rate equal to its replenishment by weathering and erosion. For degradation to continue, the thickness of valley alluvium must be slightly less than the vertical distance of flood scour. If the scouring depth is less than the alluvial thickness, the underlying bedrock cannot be attacked. If the scour activity greatly exceeds the depth of alluvium, the excess energy of flood runoff can sweep the valley clean of all debris, and the flood channels will be deeply incised in the bedrock floor. The active and abandoned channels in the valley bottom expose bedrock in many places but are never incised deeply below the bedrock surface. Thus, the depth of scour must average slightly greater than the depth of alluvium on the valley bottoms.

Though an average depth of scour will prevail under uniform climatic conditions, in-

dividual floods should vary somewhat in intensity, resulting in greater or lesser depth of scour, reflected in the bedrock surface by minor channel irregularities. This is shown in these valleys by the numerous bedrock exposures in the irregularly trending channels. In places the bedrock is incised to a few feet in depth; in other localities it is covered by a few feet of alluvial debris.

Valley deepening and valley widening occur simultaneously in this region, as revealed in variations in widths of valleys. Locally, aided by less resistant rock types or the presence of secondary rock structures, retreat of valley walls is more rapid than valley deepening, so that canyon basins and broad piedmont valley flats are being opened out. Where rock is more resistant the sidewall retreat is negligible, and the valleys remain narrow. The writer believes the valley bottoms are being slowly and uniformly lowered, and locally widened. The downcutting is accomplished by the vertical corrasion activity of sporadic runoff, shifted laterally by alluvial damming in the more open valley segments and somewhat more confined in the narrow segments. The result is a uniform lowering of the valley floor in all segments. The steepening of the gradient through narrow valley segments, where the rock is more resistant, may be due, at least in part, to the concentration of flood runoff in a more limited space. However, deep incision along any channel, no matter how confined, cannot occur without permanent flow.

Development of the Piedmont Surfaces

Slight fluctuation in precipitation conditions is the major cause of the various piedmont levels. The processes which developed these levels are now active in many of the piedmont valleys and in restricted segments of the mountain canyons. Valley widening under the present semiarid regime is at an optimum. The piedmont surfaces were formed during similar episodes of valley widening. The vertical intervals between the levels represent periods of valley deepening brought about by a change in precipitation conditions. The alternations between valley widening and valley deepening were controlled by such slight fluctuations over long periods of time.

If the climate becomes humid enough to promote continuous runoff, stream channels will be permanently confined. The vegetative cover will increase, promoting the activity of organic acids, accelerating the formation of deep soils, and thus effectively halting retreat of slopes. Stabilization decreases the amount of moving debris. The products of weathering reaching the valley bottom would be finer, readily removable by the increased runoff. Lack of factors favoring alluvial damming will inhibit channel shifting, and downward cutting along one line of flow will result.

It has been suggested that more humid conditions would initiate mass wasting of the valley sides, thus continuing the process of slope retreat. However, this writer believes that a heavier vegetative cover would slow down mass wasting and would result in a decrease in the size of debris fragments reaching the valley floor. Hence, accumulations of coarse-textured debris at the bases of sidewall slopes would be less common. Furthermore, the elimination of sudden, localized, torrential runoff, and the increased tendency for streams to maintain a constant flow in a fixed position in the valley floors, should increment downcutting. However, the elimination of braiding with the fluctuation to subhumid conditions may result in development of meanders. Local slope retreat might thus be maintained by meander swinging.

Extreme aridity will not produce the proper circumstances for valley widening. A greatly decreased runoff could not remove the coarse debris supplied by predominantly physical weathering processes. This debris, since scanty vegetation would mean relative absence of organic acids, would tend to remain fresh and coarse. Valleys would choke up, and sidewall retreat would be arrested by the cloaking effect of the debris accumulations. Furthermore, downcutting would be inhibited by the great depth of valley alluvium.

Conditions favoring valley widening require balance between the amount of debris supplied by weathering and the ability of the runoff to remove it. Semiaridity, such as exists in the region at present, appears to best satisfy these requirements. The cycle of climatic fluctuation from semiarid to slightly more humid will yield

more rapid vertical incision; a shift back to semiarid yields a predominance of valley widening. Throughout all portions of the cycle, however, the valleys are continuously deepening.

Influence of the Dakota Ridge

Elevation, trend, size, and extent of preservation of the surfaces are related to the presence or absence of the Dakota ridge. This relationship is best expressed in the Deadman Canyon section (Fig. 4), where the various levels are well enough preserved to reveal the control exerted by this unit.

The higher elevations of the High Piedmont are largely due to the confining influence of this ridge, although the fact that more resistant strata underlie the area is a strong contributing factor. Surfaces of the Low-Level group are at greater elevations in the High Piedmont than their equivalents in adjacent portions of the Low Piedmont (Fig. 2). The correlation between the two areas was accomplished in part by actual tracing of remnant levels. Additional correlative evidence is had from comparable conditions of surface dissection patterns and sequential position in the series of levels. The lower elevations of levels along Little Bear and Little Fountain creeks, as compared with their counterparts west of the Dakota ridge along Little Turkey Creek, are the consequence of more rapid downcutting in less resistant rocks (Pierre shale). Modern comparison upholds this fact. Little Bear Creek canyon is at 6550 feet at its mouth, whereas Little Turkey Creek canyon is at 6825 feet in its mouth portion (Fig. 2). The lower position of Little Bear Creek, and Little Fountain Creek as well, even though the upper piedmont segments of these streams are in the same rock types as Little Turkey Creek, can be attributed to deeper incision in the weak shales east of the Dakota ridge and headward migration of this deepening into the more resistant rocks to the west. It is notable that all streams passing through the Dakota ridge have a marked gradient increase in their ridge segment (Fig. 2).

The Dakota unit's control on the trend of surfaces is exemplified in the remnants along the north side of Little Turkey Creek (Fig. 4, (5b), (5c), (5d)). This level extends southeast for approximately a mile toward the ridge and then abruptly turns to the south. Instances of similar control by topographic obstacles occur elsewhere in the piedmont—for example, the fan-shaped remnant west of Popes Bluff in the norther portion of the quadrangle (Pl. 1).

The presence of the Dakota has quite effectively slowed dissection of the High Piedmont, acting as a local base level in this respect. It has also acted as a barrier, diverting drainage lines from the Deadman Canyon surfaces and causing the main streams to pass to the northeast and west of that remnant. Thus, the broad preservation of High-Level group surfaces in this locality is the direct result of the presence of the Dakota unit. The effect on extent of surfaces is clearly shown by comparing the areal distribution of Low-Level remnants in the High and Low piedmont areas (Pl. 1).

Erosional History of the Deadman Canyon Area[2]

The Deadman Canyon surfaces are the oldest landforms preserved in this portion of the Colorado piedmont. In former extent these High-Level group surfaces coalesced to form the floor of the piedmont. Each developed as an ancestral valley, widened laterally by sidewall retreat and with underlying bedrock surfaces smoothed by the scouring action of torrential runoff. The occurrence of equivalent remnants throughout the quadrangle (Pl. 1) indicates that similar broad valley flats of the same age were once widespread. The northern portion of the Deadman Canyon surfaces developed along an ancestral drainage line which issued from the Mountain Mass in the vicinity of Little Fountain Creek canyon (1). The presence of rhyolite debris in the surface alluvium, the up-stream convergence of old channels in the surface, and the accordant remnants in the direction of former mountainward extent support this. The Dakota ridge (X-Y) projected above this level.

The level preserved in the southern portion (2a)(2b) is a relic of a similar flat along the ancestral Little Turkey-Turkey Creek drainage lines. Levels (1) and (2) coalesced east of the Dakota ridge to form the floor of the pied-

[2] All references in this portion of the text are keyed to Figure 4.

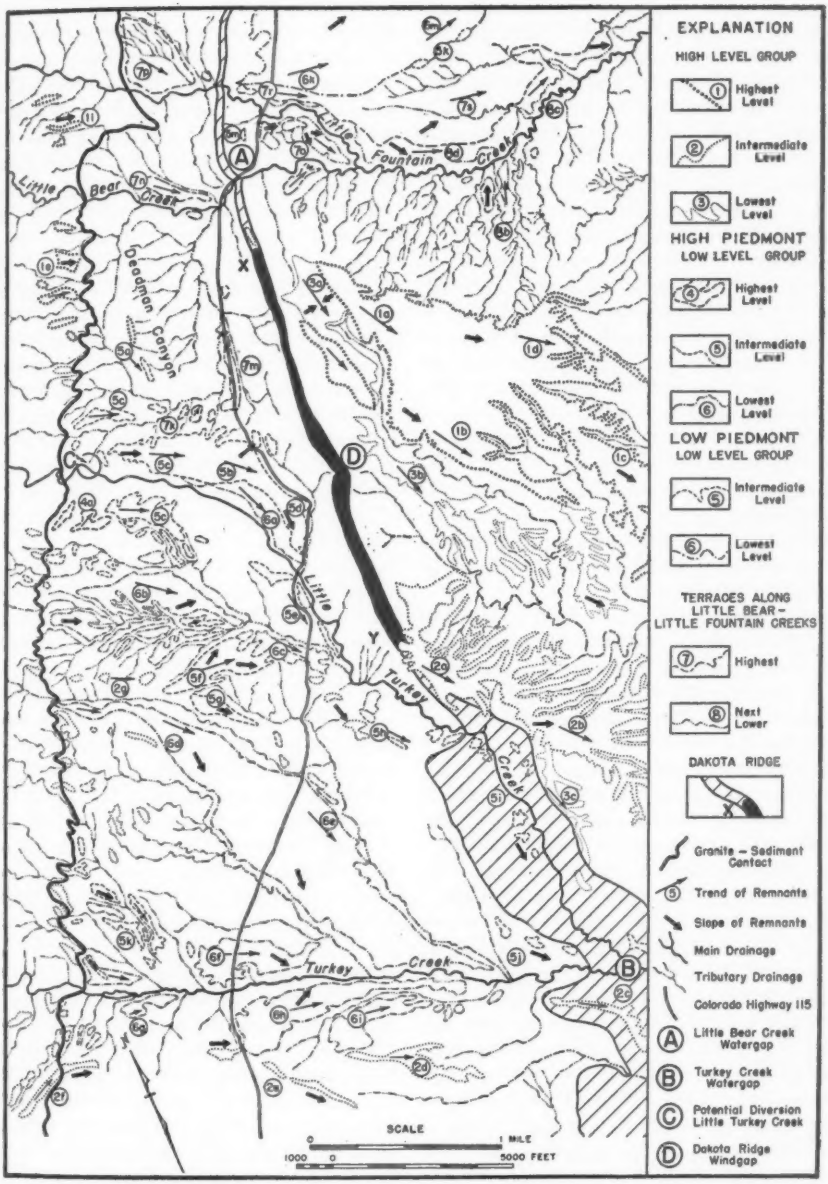

FIGURE 4.—MAP SHOWING PIEDMONT INTERSTREAM SURFACE REMNANTS IN THE DEADMAN CANYON AREA

This map is keyed to the text description of the erosional history of the Deadman Canyon area.

mont of that time. West of the ridge, however, a drainage divide separated the two valleys. The virtual absence of rhyolite debris in the southern level upholds this contention;

the ancestral Little Turkey Creek did not have access to the outcrop of this rock (1e). A wind gap in the Dakota ridge (D) at 6650 feet preserves one small segment of an old Little Turkey Creek course. Mountain-spur remnants developed by the Little Turkey Creek drainage (2g), as well as small elements in the High Piedmont along Turkey Creek, reveal the former mountainward extent of this old valley floor (2f) (2d) (2e), extending it across the Dakota outcrop through the Turkey Creek water gap (2c). The surface beveled the Dakota for a considerable distance south of (Y), although the unit continued farther south as a ridge along the western edge of the valley. The widespread coalescence of High-Level group surfaces to form the early piedmont floor is suggested by remnants 30 or more miles from the mountain front, near the Fountain Creek-Arkansas River junction. Here the levels stand 200 feet above the present valley bottoms. Doubtless, many streams contributed to form this old piedmont floor.

The maximum alluvial thicknesses in the Deadman Canyon surfaces, 110 and 40 feet, seem to require a depth of scour greater than that characteristic of the modern valley bottoms. However, data from comparable climatic regimes reveal that even a depth of scour of 100 feet is not unreasonable (Bailey, 1934). Furthermore, the roughly planate character of the underlying bedrock surface, along with the presence of water-worn alluvium directly overlying that surface, indicates development by running water. The location of these excessive alluvial sections suggests that they may represent valley-profile segments where steep-gradient streams encountered reduced gradients in the piedmont below the mountain base. Thicker deposits may be expectable under such circumstances. The upslope and downslope lensing character of the deposits and the semblance of bedding maintained throughout indicate gradual accretion rather than a sudden accumulation during one short episode. No "valley-choking" effect resulting from aridity is suggested. The alluvial debris of these thicker sections was derived from the Mountain Mass and transported by water to its present position in the piedmont. With greater aridity, thick alluvial sections should also be present adjacent to and within the mountain area. Such is not the case. A possible relationship has been suggested between glaciation in the higher mountain sections and alluviation in the piedmont. This proposal is reserved for future investigation. However, climatic change favorable for or adverse to glaciation may correlate with increase or decrease of aridity in the region.

During the late stages of alluviation of the highest piedmont flats, the runoff shifted about on the surface to produce the braided channel patterns. Eventually, however, perhaps as a result of more humid conditions, the stream in the last-occupied channel was incised. A short episode of valley widening ensued, and lower, more restricted flats developed below the older levels. Remnants of this lower surface are preserved along the eastern base of the Dakota ridge (3a)(3b) and along the east side of Turkey Creek (3c).

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Since the development of this lowest piedmont interstream surface, alternating episodes of valley deepening and valley widening have produced terraces in the walls of the modern valleys. In the Low Piedmont along Little Fountain Creek, the two highest terraces are widespread; the higher one approaches the dimensions of a piedmont interstream surface (7r) (7o) (7s). This highest terrace is also preserved upstream along the Little Bear-Little Fountain Creek drainage west of the Dakota ridge (7p) (7n) (7m) (7k). The next lower terrace (8d) (8c) is somewhat narrower but nevertheless quite broad. The remnant marked (8b) is a tributary flat to this terrace level. Fragments of two lower terraces, not shown in Figure 4, are also present along this drainage. Farther east, several additional minor terrace levels indicate local valley deepening and widening which did not progress far enough headward to be represented in this portion of the valley. Terrace levels in the High Piedmont valleys are few, restricted in width, and poorly developed. This can only be the result of the greater resistance of the rocks to the processes of valley deepening and valley widening.

Drainage diversions have been noted. Other diversions occur or are about to occur in many localities. One such potential point of capture is along Little Turkey Creek (C) where a Little Bear Creek tributary is encroaching into the Little Turkey Creek drainage area. A low, relatively narrow divide separates the two drainage lines. The smaller stream, aided by a very steep gradient, is the aggressor. Only a slight reduction of the drainage divide is required before torrential flooding of the Little Turkey Creek valley will overtop it. The erosive power of such a flood will deepen the smaller stream channel, thereby increasing the chances for repetition of the event. Eventually, Little Turkey Creek above the point of diversion will become part of the Little Bear Creek drainage area. Diversions of this sort may be responsible, at least in part, for local

variations in gradient, slope direction, elevation, and dimensions of the piedmont interstream surfaces.

Age Considerations

Finlay (1916), p. 11) concluded that the presence of a molar tooth of *Elephas columbi* in the basal portion of the Mesa surface alluvium fixes the age of the:

> "lowest (youngest) mesa gravel as Pleistocene and therefore contemporaneous with the glacial deposits in the higher valleys on Pikes Peak . . . The higher mesa gravels are in part equivalent in age to the Nussbaum gravels, mapped in the Pueblo folio, which are regarded as of late Tertiary age."

Finlay further established the age of the "tertiary gravels" around Colorado Springs as Late Pleistocene, though he did not note the occurrence of these deposits in other localities in the quadrangle.

Gilbert (1897, p. 3) made the following notation on the Nussbaum formation which he considered Neocene:

> "The Nussbaum is not conformable with any underlying formation, but rests on eroded surfaces of the Pierre, Niobrara, and other Cretaceous strata. The largest and thickest body is on Baculite Mesa. A second body occupies the mesa north of Blue Hill, and there are others between that and Turkey Creek."

Gilbert placed other alluvial deposits, occurring in steplike arrangement in all the principal valleys of the Pueblo quadrangle, in earlier and later Pleistocene, observing that:

> "Each terrace was, in fact, once the flood-plain or bottom land of the stream, and their arrangement in steps is a record of the gradual deepening of the valleys by erosion. Their order of position is also their order of age, but in an inverse way, as the lowest is the latest."

The Deadman Canyon surfaces were traced into the Pueblo quadrangle to the south where they are underlain by the fanglomerates described by Gilbert as the Nussbaum formation (see above). The highest terraces, regarded by Gilbert as early Pleistocene, are down-valley continuations of the Low-Level group of the present paper. Lower terraces, late Pleistocene in Gilbert's estimation, continue northward from Pueblo along Fountain Creek to form the terraces in the modern stream valleys of the Colorado Springs quadrangle.

Van Tuyl and Lovering (1935, p. 1349) referred to the Deadman Canyon surfaces as the "Orodell berm", which they considered Pliocene:

> "The gravel-strewn rock benches below the high-level gravels associated with the Orodell surface are closely related to the present valleys. The character of the gravels, as well as the interval between the benches on which they lie, suggests a correlation with rock benches and terrace gravels farther west in the mountains where the relationship to various Pleistocene glacial epochs is obvious."

These authors also state:

> "It is certain that at least two stages of Pleistocene glaciation occurred in the Front Range, and the intermediate terraces suggest a third . . . the valleys were deepened in successive glacial stages. As the climate became more humid the volume of the streams increased, erosion became more marked, and canyon-cutting ensued. As the climate became drier during the interglacial periods, valley-deepening ceased, and the streams spent their energy in lateral corrasion and transportation, while the interstream areas were being lowered by sheetwash and pedimentation. The terrace gravels may, in large part, mark the periods of ice recession and the related interglacial stages. The climatic changes of the Pleistocene seem adequate to explain the successive terraces, but, in addition, regional uplifts may be involved."

The age determinations of Finlay, Gilbert, or Van Tuyl and Lovering are not conclusive. The occurrence of *Elephas columbi*, or other evidence cited, has little weight, other than to suggest that all gravels and interstream surfaces of this region are comparatively young.

Wahlstrom (1947, p. 568) regarded the differentiation of levels as related to mountain uplift:

> "The uplift of the Front Range to its present elevation was not the result of a single upheaval. The presence of more or less poorly developed terraces in the mountains and very well developed terraces in the valleys east of the mountains suggests intermittent uplifts. The terraces are cut below and postdate the early Pleistocene ice-sheet till."

While the postulate of intermittent uplifts may explain different erosion levels, it is not in accord with diastrophic probability. This writer knows of no orogenic quality that would produce halts and starts in the uplift of so vast a region as the Front Ranges. More probably the differentiation of the levels is controlled externally by slight alterations of the climate

in cyclic fashion, and mountain uplift has been relatively constant. However, the piedmont interstream surfaces do post-date early Pleistocene.

The oldest High-Level surfaces are distinctly younger than the so-called Rocky Mountain peneplain. The latter is at an elevation of 9500–10,000 feet on the mountain tops along the western edge of the piedmont. The highest piedmont surface has a mountain-front elevation of 7500 feet. Wahlstrom assigned a Pliocene or early Pleistocene age to the Rocky Mountain peneplain. Gilbert, by dating the Nussbaum gravels as Pliocene, would have the highest piedmont interstream surface as Pliocene. Obviously he was in error. The difference in elevation between the Rocky Mountain peneplain and this highest piedmont surface suggests a considerable time interval between them. Furthermore, the Deadman Canyon surfaces could not correlate with Van Tuyl and Lovering's Orodell berm, considered Pliocene. It appears most likely that the High-Level group surfaces are Pleistocene, probably middle Pleistocene, and the Low-Level group surfaces are later Pleistocene and Recent. The still younger stream terraces are certainly Recent.

Conclusions

The piedmont interstream surfaces of the Colorado Springs region were developed by a combination of sidewall retreat and stream scour such as is occurring today. Within the Mountain Mass these processes are forming localized basins below steep canyon walls. In the Low Piedmont, underlain by rocks of weak resistance, wide valley flats are developing; in the High Piedmont, underlain by more resistant rocks, the extent of valley bottoms is more limited. These widen downvalley, especially across the outcrop of the Pierre shale in the Low Piedmont. The sole difference between the small canyon basins, the restricted valleys of the High Piedmont, and the broad, flaring valleys of the Low Piedmont is in size.

Semiarid climate prevails, and the region is subjected to occasional, local, heavy precipitation accompanied by flood runoff. Torrential stream flow is sufficient to remove the debris of weathering from the base of the slopes. Streams are able to scour their channels locally into the underlying bedrock beneath the valley alluvium. Alluvial obstacles divert the flow into various positions on the valley floor so that all portions experience bedrock scour. Since the depth of scour is comparatively uniform, with long-continued uniformity of climatic conditions, a relatively flat bedrock floor is developed beneath the alluvium. Where channels are forced against valley walls, local widening also occurs by lateral corrasion. Sidewall retreat results largely from weathering of the exposed rock but is aided by lateral corrasion, and perhaps by "gully gravure" (Bryan, 1940a), particularly in the weak rocks of the Low Piedmont.

The differentiation of piedmont surfaces into various levels appears to relate to climatic fluctuations, rather than the intermittent uplift suggested by Wahlstrom (1947). Under semiarid climatic conditions, presumably during the drier interglacial epochs, canyon-basin development and valley widening are promoted. During more humid intervals, the streams cut their channels lower at faster rates; the increase in precipitation perhaps correlates with the incidence of glaciation in higher mountain areas. Valley widening through such episodes is subordinate. An alternation between subhumid and semirarid climate would account for the differentiation of piedmont levels, even as mountain uplift was uniform and continuous. This thesis agrees more closely with the known facts of Pleistocene climatic change and accepted theories of mountain development than does the postulate of intermittent uplifts.

The oldest, highest piedmont interstream surface developed between valley walls. An uparched flank of the mountain region extended considerably eastward from the position of the mountain front today. Canyons were cut in the granites and other resistant rocks and broad flats opened out in the less-resistant sediments to the east and southeast. This interval of valley widening was long, and the piedmont surfaces to the east became very widespread, coalescing in their lower portions to form the floor of the piedmont. The author believes this occurred in middle or late middle Pleistocene time. With continuation of mountain uparching, but under more humid conditions, canyons were deepened and piedmont

surfaces dissected and uplifted. A return to semiarid conditions brought on a new episode of valley widening at a lower level. Broad surfaces were again developed, widely distributed in the Low Piedmont, more restricted in the High Piedmont. A shift back to more humid conditions followed, and the surfaces were dissected as streams cut downward. In the modern scene, valley widening is prevalent as the climatic regime shifts toward the semiarid. Three main episodes of valley widening have occurred: that producing the High-Level surfaces; that producing the Low-Level group; and that taking place today. Minor fluctuations throughout the cycles account for the lesser levels within each major group and for the stream terraces in the modern valleys.

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UPPER CAMBRIAN STRATIGRAPHY IN THE CENTRAL APPALACHIANS

BY JAMES LEE WILSON

ABSTRACT

This paper presents a stratigraphic and paleontologic study of the Upper Cambrian exposed in three linear belts in the central Appalachians: (1) in the Nittany Arch region of southern Pennsylvania (Warrior and Gatesburg formations), (2) in the Shenandoah-Cumberland Valley of Virginia, West Virginia, Maryland, and Pennsylvania (Elbrook and Conococheague formations), and (3) the Frederick Valley (Frederick limestone) of Maryland. Cambrian stratigraphy in the three areas is discussed, and a new member is named in the lower part of the Conococheague formation. A general facies change west to east from arenaceous dolomite (Gatesburg formation) to silt-laminated limestone (Conococheague formation) is indicated in the geosyncline.

Charles Butts' earlier correlation of the Warrior with the Elbrook and the Gatesburg with the Conococheague formation is found to be approximately correct on the basis of trilobite faunas. The paleogeographic significance of the stratigraphy is outlined on a regional isopachous-facies map.

Trilobites identified from the Warrior, upper Elbrook, and lowest Conococheague formations indicate a Dresbachian age for these beds. Trilobites of Franconian age from the middle Gatesburg and lower Conococheague formations show more complex relationships. The Ore Hill member of the Gatesburg formation contains in descending order the *Conaspis* and *Elvinia* faunizones, the latter replaced eastward in the Conococheague formation by a new trilobite facies fauna. The fauna is named for its most abundant trilobite, *Pseudosaratogia magna*, and in the Conococheague constitutes a lower Franconian faunizone equivalent to the *Elvinia* unit in the standard sequence. Trempealeauian trilobites are identified from the upper Conococheague formation.

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the ancestral Little Turkey Creek did not have access to the outcrop of this rock (1e). A wind gap in the Dakota ridge (D) at 6650 feet preserves one small segment of an old Little Turkey Creek course. Mountain-spur remnants developed by the Little Turkey Creek drainage (2g), as well as small elements in the High Piedmont along Turkey Creek, reveal the former mountainward extent of this old valley floor (2f) (2d) (2e), extending it across the Dakota outcrop through the Turkey Creek water gap (2c). The surface beveled the Dakota for a considerable distance south of (Y), although the unit continued farther south as a ridge along the western edge of the valley. The widespread coalescence of High-Level group surfaces to form the early piedmont floor is suggested by remnants 30 or more miles from the mountain front, near the Fountain Creek-Arkansas River junction. Here the levels stand 200 feet above the present valley bottoms. Doubtless, many streams contributed to form this old piedmont floor.

The maximum alluvial thicknesses in the Deadman Canyon surfaces, 110 and 40 feet, seem to require a depth of scour greater than that characteristic of the modern valley bottoms. However, data from comparable climatic regimes reveal that even a depth of scour of 100 feet is not unreasonable (Bailey, 1934). Furthermore, the roughly planate character of the underlying bedrock surface, along with the presence of water-worn alluvium directly overlying that surface, indicates development by running water. The location of these excessive alluvial sections suggests that they may represent valley-profile segments where steep-gradient streams encountered reduced gradients in the piedmont below the mountain base. Thicker deposits may be expectable under such circumstances. The upslope and downslope lensing character of the deposits and the semblance of bedding maintained throughout indicate gradual accretion rather than a sudden accumulation during one short episode. No "valley-choking" effect resulting from aridity is suggested. The alluvial debris of these thicker sections was derived from the Mountain Mass and transported by water to its present position in the piedmont. With greater aridity, thick alluvial sections should also be present adjacent to and within the mountain area. Such is not the case. A possible relationship has been suggested between glaciation in the higher mountain sections and alluviation in the piedmont. This proposal is reserved for future investigation. However, climatic change favorable for or adverse to glaciation may correlate with increase or decrease of aridity in the region.

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They were then dissected during downcutting, and eventually, at lower elevations, less extensive surfaces were produced in the High Piedmont, and more extensive surfaces were developed in the Low Piedmont. Remnants of this level, the lowest of the Low-Level group, are preserved along Turkey and Little Turkey creeks (6a) (6b) (6c) (6d) (6e) (6f) (6g) (6h) (6i). Rhyolite is scarce in the alluvium of these remnants in the High Piedmont, which is largely reworked materials from higher (older) surfaces. By this time, the headward portion of Little Turkey Creek had been diverted to the east by encroachment of the Little Bear Creek drainage, with a southward migration of the drainage divide between the two. The lower elevation and steeper gradient of the latter stream gave it a comparatively greater erosional ability. This encroachment continues. The equivalent level along Little Bear and Little Fountain creeks extends in fanshaped form eastward from the water gaps of these streams through the Dakota ridge into the Low Piedmont (6k) (6m).

Since the development of this lowest piedmont interstream surface, alternating episodes of valley deepening and valley widening have produced terraces in the walls of the modern valleys. In the Low Piedmont along Little Fountain Creek, the two highest terraces are widespread; the higher one approaches the dimensions of a piedmont interstream surface (7r) (7o) (7s). This highest terrace is also preserved upstream along the Little Bear-Little Fountain Creek drainage west of the Dakota ridge (7p) (7n) (7m) (7k). The next lower terrace (8d) (8c) is somewhat narrower but nevertheless quite broad. The remnant marked (8b) is a tributary flat to this terrace level. Fragments of two lower terraces, not shown in Figure 4, are also present along this drainage. Farther east, several additional minor terrace levels indicate local valley deepening and widening which did not progress far enough headward to be represented in this portion of the valley. Terrace levels in the High Piedmont valleys are few, restricted in width, and poorly developed. This can only be the result of the greater resistance of the rocks to the processes of valley deepening and valley widening.

Drainage diversions have been noted. Other diversions occur or are about to occur in many localities. One such potential point of capture is along Little Turkey Creek (C) where a Little Bear Creek tributary is encroaching into the Little Turkey Creek drainage area. A low, relatively narrow divide separates the two drainage lines. The smaller stream, aided by a very steep gradient, is the aggressor. Only a slight reduction of the drainage divide is required before torrential flooding of the Little Turkey Creek valley will overtop it. The erosive power of such a flood will deepen the smaller stream channel, thereby increasing the chances for repetition of the event. Eventually, Little Turkey Creek above the point of diversion will become part of the Little Bear Creek drainage area. Diversions of this sort may be responsible, at least in part, for local

variations in gradient, slope direction, elevation, and dimensions of the piedmont interstream surfaces.

Age Considerations

Finlay (1916), p. 11) concluded that the presence of a molar tooth of *Elephas columbi* in the basal portion of the Mesa surface alluvium fixes the age of the:

"lowest (youngest) mesa gravel as Pleistocene and therefore contemporaneous with the glacial deposits in the higher valleys on Pikes Peak . . . The higher mesa gravels are in part equivalent in age to the Nussbaum gravels, mapped in the Pueblo folio, which are regarded as of late Tertiary age."

Finlay further established the age of the "tertiary gravels" around Colorado Springs as Late Pleistocene, though he did not note the occurrence of these deposits in other localities in the quadrangle.

Gilbert (1897, p. 3) made the following notation on the Nussbaum formation which he considered Neocene:

"The Nussbaum is not conformable with any underlying formation, but rests on eroded surfaces of the Pierre, Niobrara, and other Cretaceous strata. The largest and thickest body is on Baculite Mesa. A second body occupies the mesa north of Blue Hill, and there are others between that and Turkey Creek."

Gilbert placed other alluvial deposits, occurring in steplike arrangement in all the principal valleys of the Pueblo quadrangle, in earlier and later Pleistocene, observing that:

"Each terrace was, in fact, once the flood-plain or bottom land of the stream, and their arrangement in steps is a record of the gradual deepening of the valleys by erosion. Their order of position is also their order of age, but in an inverse way, as the lowest is the latest."

The Deadman Canyon surfaces were traced into the Pueblo quadrangle to the south where they are underlain by the fanglomerates described by Gilbert as the Nussbaum formation (see above). The highest terraces, regarded by Gilbert as early Pleistocene, are down-valley continuations of the Low-Level group of the present paper. Lower terraces, late Pleistocene in Gilbert's estimation, continue northward from Pueblo along Fountain Creek to form the terraces in the modern stream valleys of the Colorado Springs quadrangle.

Van Tuyl and Lovering (1935, p. 1349) referred to the Deadman Canyon surfaces as the "Orodell berm", which they considered Pliocene:

"The gravel-strewn rock benches below the high-level gravels associated with the Orodell surface are closely related to the present valleys. The character of the gravels, as well as the interval between the benches on which they lie, suggests a correlation with rock benches and terrace gravels farther west in the mountains where the relationship to various Pleistocene glacial epochs is obvious."

These authors also state:

"It is certain that at least two stages of Pleistocene glaciation occurred in the Front Range, and the intermediate terraces suggest a third . . . the valleys were deepened in successive glacial stages. As the climate became more humid the volume of the streams increased, erosion became more marked, and canyon-cutting ensued. As the climate became drier during the interglacial periods, valley-deepening ceased, and the streams spent their energy in lateral corrasion and transportation, while the interstream areas were being lowered by sheetwash and pedimentation. The terrace gravels may, in large part, mark the periods of ice recession and the related interglacial stages. The climatic changes of the Pleistocene seem adequate to explain the successive terraces, but, in addition, regional uplifts may be involved."

The age determinations of Finlay, Gilbert, or Van Tuyl and Lovering are not conclusive. The occurrence of *Elephas columbi*, or other evidence cited, has little weight, other than to suggest that all gravels and interstream surfaces of this region are comparatively young.

Wahlstrom (1947, p. 568) regarded the differentiation of levels as related to mountain uplift:

"The uplift of the Front Range to its present elevation was not the result of a single upheaval. The presence of more or less poorly developed terraces in the mountains and very well developed terraces in the valleys east of the mountains suggests intermittent uplifts. The terraces are cut below and postdate the early Pleistocene ice-sheet till."

While the postulate of intermittent uplifts may explain different erosion levels, it is not in accord with diastrophic probability. This writer knows of no orogenic quality that would produce halts and starts in the uplift of so vast a region as the Front Ranges. More probably the differentiation of the levels is controlled externally by slight alterations of the climate

in cyclic fashion, and mountain uplift has been relatively constant. However, the piedmont interstream surfaces do post-date early Pleistocene.

The oldest High-Level surfaces are distinctly younger than the so-called Rocky Mountain peneplain. The latter is at an elevation of 9500–10,000 feet on the mountain tops along the western edge of the piedmont. The highest piedmont surface has a mountain-front elevation of 7500 feet. Wahlstrom assigned a Pliocene or early Pleistocene age to the Rocky Mountain peneplain. Gilbert, by dating the Nussbaum gravels as Pliocene, would have the highest piedmont interstream surface as Pliocene. Obviously he was in error. The difference in elevation between the Rocky Mountain peneplain and this highest piedmont surface suggests a considerable time interval between them. Furthermore, the Deadman Canyon surfaces could not correlate with Van Tuyl and Lovering's Orodell berm, considered Pliocene. It appears most likely that the High-Level group surfaces are Pleistocene, probably middle Pleistocene, and the Low-Level group surfaces are later Pleistocene and Recent. The still younger stream terraces are certainly Recent.

Conclusions

The piedmont interstream surfaces of the Colorado Springs region were developed by a combination of sidewall retreat and stream scour such as is occurring today. Within the Mountain Mass these processes are forming localized basins below steep canyon walls. In the Low Piedmont, underlain by rocks of weak resistance, wide valley flats are developing; in the High Piedmont, underlain by more resistant rocks, the extent of valley bottoms is more limited. These widen downvalley, especially across the outcrop of the Pierre shale in the Low Piedmont. The sole difference between the small canyon basins, the restricted valleys of the High Piedmont, and the broad, flaring valleys of the Low Piedmont is in size.

Semiarid climate prevails, and the region is subjected to occasional, local, heavy precipitation accompanied by flood runoff. Torrential stream flow is sufficient to remove the debris of weathering from the base of the slopes. Streams are able to scour their channels locally into the underlying bedrock beneath the valley alluvium. Alluvial obstacles divert the flow into various positions on the valley floor so that all portions experience bedrock scour. Since the depth of scour is comparatively uniform, with long-continued uniformity of climatic conditions, a relatively flat bedrock floor is developed beneath the alluvium. Where channels are forced against valley walls, local widening also occurs by lateral corrasion. Sidewall retreat results largely from weathering of the exposed rock but is aided by lateral corrasion, and perhaps by "gully gravure" (Bryan, 1940a), particularly in the weak rocks of the Low Piedmont.

The differentiation of piedmont surfaces into various levels appears to relate to climatic fluctuations, rather than the intermittent uplift suggested by Wahlstrom (1947). Under semiarid climatic conditions, presumably during the drier interglacial epochs, canyon-basin development and valley widening are promoted. During more humid intervals, the streams cut their channels lower at faster rates; the increase in precipitation perhaps correlates with the incidence of glaciation in higher mountain areas. Valley widening through such episodes is subordinate. An alternation between subhumid and semirarid climate would account for the differentiation of piedmont levels, even as mountain uplift was uniform and continuous. This thesis agrees more closely with the known facts of Pleistocene climatic change and accepted theories of mountain development than does the postulate of intermittent uplifts.

The oldest, highest piedmont interstream surface developed between valley walls. An uparched flank of the mountain region extended considerably eastward from the position of the mountain front today. Canyons were cut in the granites and other resistant rocks and broad flats opened out in the less-resistant sediments to the east and southeast. This interval of valley widening was long, and the piedmont surfaces to the east became very widespread, coalescing in their lower portions to form the floor of the piedmont. The author believes this occurred in middle or late middle Pleistocene time. With continuation of mountain uparching, but under more humid conditions, canyons were deepened and piedmont

surfaces dissected and uplifted. A return to semiarid conditions brought on a new episode of valley widening at a lower level. Broad surfaces were again developed, widely distributed in the Low Piedmont, more restricted in the High Piedmont. A shift back to more humid conditions followed, and the surfaces were dissected as streams cut downward. In the modern scene, valley widening is prevalent as the climatic regime shifts toward the semiarid. Three main episodes of valley widening have occurred: that producing the High-Level surfaces; that producing the Low-Level group; and that taking place today. Minor fluctuations throughout the cycles account for the lesser levels within each major group and for the stream terraces in the modern valleys.

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UPPER CAMBRIAN STRATIGRAPHY IN THE CENTRAL APPALACHIANS

BY JAMES LEE WILSON

ABSTRACT

This paper presents a stratigraphic and paleontologic study of the Upper Cambrian exposed in three linear belts in the central Appalachians: (1) in the Nittany Arch region of southern Pennsylvania (Warrior and Gatesburg formations), (2) in the Shenandoah-Cumberland Valley of Virginia, West Virginia, Maryland, and Pennsylvania (Elbrook and Conococheague formations), and (3) the Frederick Valley (Frederick limestone) of Maryland. Cambrian stratigraphy in the three areas is discussed, and a new member is named in the lower part of the Conococheague formation. A general facies change west to east from arenaceous dolomite (Gatesburg formation) to silt-laminated limestone (Conococheague formation) is indicated in the geosyncline.

Charles Butts' earlier correlation of the Warrior with the Elbrook and the Gatesburg with the Conococheague formation is found to be approximately correct on the basis of trilobite faunas. The paleogeographic significance of the stratigraphy is outlined on a regional isopachous-facies map.

Trilobites identified from the Warrior, upper Elbrook, and lowest Conococheague formations indicate a Dresbachian age for these beds. Trilobites of Franconian age from the middle Gatesburg and lower Conococheague formations show more complex relationships. The Ore Hill member of the Gatesburg formation contains in descending order the *Conaspis* and *Elvinia* faunizones, the latter replaced eastward in the Conococheague formation by a new trilobite facies fauna. The fauna is named for its most abundant trilobite, *Pseudosaratogia magna*, and in the Conococheague constitutes a lower Franconian faunizone equivalent to the *Elvinia* unit in the standard sequence. Trempealeauian trilobites are identified from the upper Conococheague formation.

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Introduction

Purpose of the Study

Upper Cambrian formations in the central Appalachians crop out in three linear belts: (1) along the complex axis of the Nittany Arch, the westernmost great fold in the Appalachians through central Pennsylvania; (2) along axes of two anticlines within the Cumberland-Shenandoah Valley west of South Mountain anticlinorium (southeastern Pennsylvania, Maryland and West Virginia panhandles, and northern Virginia); and (3) east of South Mountain in the graben of the Frederick Valley, central Maryland (Fig. 1). These belts trend roughly north-northeast approximately 40 miles from each other along a southeast line from Bedford, Pennsylvania, through Hagerstown, Maryland, to Frederick, Maryland. Different formations have been described in each of these belts by Butts (1918), Stose (1906), and Bassler (1919). These have been considered time-stratigraphic equivalents (Butts, 1940, p. 90; 1945, p. 3; Resser, 1938, p. 22; Stose and Stose, 1946, p. 51), but no detailed paleontology has been ever undertaken except a slight amount by Walcott and Resser (1938). The present study, therefore, redescribes the formations, identifies their faunas, and checks their presumed correlation. The discovery of a new trilobite fauna in the lower Conococheague formation of the Shenandoah-Cumberland Valley belt (Wilson, 1951) has aided detailed correlation as has the description of the trilobite faunas of the Ore Hill member of the Gatesburg formation in the western Nittany Arch belt (Wilson, 1951). Some ideas are presented about Upper Cambrian facies change and paleogeography in the central Appalachians.

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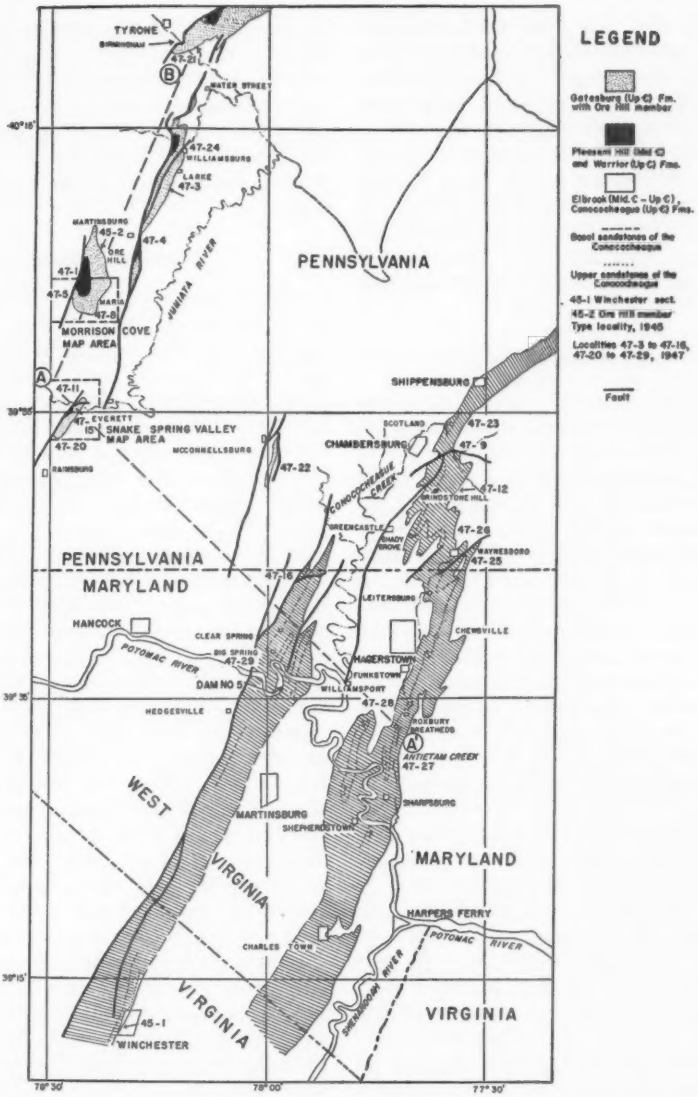


FIGURE 1.—LOCALITY INDEX MAP

Compiled from: Butts, 1946; Butts and Edmundson ms.; E. Cloos, 1941; Grimsley, 1916; Stose, 1909; 1931; ms. Revision and additions by J. L. Wilson. Scale: 1 inch = 20 miles.

FORMATIONS IN THE NITTANY ARCH BELT

General Description

Approximately 4000 feet of Cambrian beds is exposed along the Nittany Arch, brought to the surface in large part by reverse faulting along its crest. The oldest exposed beds are coarse sandstones and red and green shales correlated with the Lower Cambrian Waynesboro formation of the Cumberland Valley.

Overlying these beds is the Middle Cambrian Pleasant Hill limestone followed by the Upper Cambrian Warrior and Gatesburg formations. The area examined along this belt lies in south-central Pennsylvania largely in the following quadrangles from north to south: Bellefonte, Tyrone, Huntingdon, Hollidaysburg, Everett, and Clearville. Detailed geologic maps exist for the first four of these (Butts and Moore, 1936; Butts *et al.*, 1939; Butts, 1945), and the writer has mapped the Cambrian outcrops southward through the last two (Figs. 2, 3).

Pleasant Hill Formation

This Middle Cambrian formation was named by Butts (1918) for exposures at Pleasant Hill Church 1 mile northwest of Henrietta, Blair County, Pennsylvania. It is discussed in the Tyrone quadrangle report (Butts *et al.*, 1939), but the most detailed description is given in the Huntingdon-Hollidaysburg folio (Butts, 1945, p. 2). A thickness of 600 feet is noted at the type locality where the unit is divided into a lower two-thirds of "thin layered argillaceous, sandy, and micaceous limestone with some calcareous shale" and an upper third of "thick-bedded, fine-grained dark-gray limestone". The writer measured the upper 350 feet of the formation at the Williamsburg, Pennsylvania, section (47-24w; Butts, 1945). The upper 190 feet of the Pleasant Hill beds resemble those at the type locality in their mottled, worm-eaten appearance. They are much purer limestones than those lower in the formation. Unlike the massive fresh limestones at Pleasant Hill Church they weather to brownish-yellow thin-bedded platy layers up to 6 inches thick. Contrary to Butts' belief (1945, p. 2) there are sandy layers and some siltstone and shale quite high in the formation. A new exposure of the Pleasant Hill exists on the south side of the river 1 mile west of Williamsburg in a road cut directly east of the reverse fault described in the Williamsburg section (47-24w).

Middle Cambrian genera (*Helcionella*, *Nisusia*, *Alokistocare*, and *Kootenia*) have been collected from the upper beds of the formation at the type locality, and a *Kootenia* pygidium was obtained about 90 feet below the top of the Pleasant Hill limestone at Williamsburg. Resser has correlated this fauna with the Rogersville shale of Tennessee; he considers it younger than the single known Middle Cambrian fauna of the Elbrook which is supposedly of Rutledge age (Resser, 1938, p. 21, 22). The Pleasant Hill fauna is probably equivalent to the *Glossopleura-Kootenia* zone and low in the Middle Cambrian.

Warrior Formation

General description.—The Warrior "limestone" was named (Butts, 1918, p. 528) from exposures on "Warrior Creek in the northern part of Huntingdon County, east of Warrior's Mark, Pennsylvania". The type locality was later described in more detail (Butts *et al.*, 1939, p. 9) and has been recently studied by Tasch (1951, p. 282). One of the best and most nearly complete sections of the Warrior formation, at Williamsburg, Pennsylvania, was measured by the writer (47-24w) who found it to be 1340 feet thick (about 90 feet thicker than did Butts, 1945, p. 2). Another thick section of the formation was measured by the writer 5 miles west of State College, Pennsylvania, on U. S. Route 322 near Waddle Road junction (45-3w). The upper 750 feet of this section has been remeasured by Tasch (1951, p. 276–278). Tasch has also carefully collected and described the fauna of this section. The present writer includes an additional 200 feet of lower beds in this section, extending it down to the Old Quarry although Tasch implies that these may be higher beds below a reverse fault. The writer finds no field evidence for the postulated fault.

Two sections of the top 300 feet of the Warrior have been studied and their fauna collected by the writer in the Everett quadrangle, one at Schneider Farm (47-1w), Morrison Cove, and the other on the Pennsylvania Turnpike (47-11w) in Friend's Cove about 50 miles south of the type locality. Thickness of the Warrior formation exposed in Friend's Cove has been given as about 1200 feet (Tasch, 1951, p. 279), but no more than 300 feet of the upper beds can be seen. Tasch's quotation of Swartz discussion of the Warrior in "Friend's Cove" (1951, p. 279) refers to the Williamsburg section about 30 miles farther north in Morrison Cove.

Lithologic detail.—The Warrior formation

possesses several rock types; the most common is dark platy to medium-bedded, fine-grained to aphanitic limestone generally oölitic and fossiliferous. Accompanying these are algal stromatolites of the type *Cryptozoon undulatum* Bassler 1919, and interbedded with them is dark aphanitic silty rusty-weathering dolomite. Tasch has discussed the cyclical deposition and paleoecology of these beds (1951, p. 284, 285). The dark platy limestone occurs at the top of the formation southward to the Turnpike section and is more resistant and better exposed than other beds in the Warrior. The Williamsburg section (47-24w) contains massive dolomite of the Gatesburg type, much of it dark and coarse to medium crystalline, although light dolomite also occurs. Butts (1945, p. 2) has described the orthoquartzite beds of the Warrior in some detail; those at Williamsburg are rather pure. The total section at Williamsburg shows about 85 per cent dolomite, 10 per cent limestone, and 5 per cent orthoquartzite.

Geographic expression and variation along strike.—The Warrior formation is less resistant than the overlying Gatesburg beds and forms rolling grass-covered topography along its uncommon outcrop belts. Except in the above-mentioned sections, it is poorly exposed, and only the limestone beds crop out. Numerous small quarries along strike expose 15 or 20 feet of this rock.

Limited field work and incomplete exposures confined to a north-south line have made impractical any subdivision of the Warrior formation. Except for the Waddle Road section near State College (45-3w), the writer has studied the formation only from Williamsburg, Pennsylvania, south to the Turnpike. The Waddle Road section is 925 feet thick and exposes the upper two-thirds of the formation. It contains more limestone and much less massive crystalline dolomite than the Williamsburg section and lacks orthoquartzite beds. The two southern sections in the Everett quadrangle show mostly limestone with interbedded dolomite. These and the more nearly complete section above indicate that there is more limestone in the top of the formation than elsewhere but no name is given to this as yet poorly known member.

Stratigraphic boundaries.—The Warrior is considered conformable with both underlying and overlying beds. At Williamsburg (47-24w) a section of vertical beds is exposed in a draw draining the north bank of the Frankstown branch of the Juniata River at its abrupt bend beneath exposures of the Pleasant Hill formation. These beds begin with impure limestone and siltstone of the Pleasant Hill and grade into purer dolomite through a thickness of 100 feet. The dolomite may be traced up the slope into the lowermost outcrops of the Warrior formation along the crest of the bluff.

In the Everett quadrangle in a road cut on Pennsylvania State Highway 868, a quarter of a mile north of where Potter Creek leaves Morrison Cove, the upper 70 feet of Warrior limestones is exposed; it grades upward into the Stacy dolomite member of the Gatesburg formation. In most sections, however, a covered interval of 50 to 300 feet separates the highest exposed Warrior limestone and the lowest exposed beds of the Gatesburg which are usually sandstone. The "hiatus of considerable magnitude" between the Warrior and Gatesburg formations (Butts *et al.*, 1939, p. 10; Butts and Moore, 1935, p. 15) is not based on field evidence but on correlations made by E. O. Ulrich when faunas of these Cambrian beds had not been adequately studied. Evidence in this report under faunal discussion of the Gatesburg formation indicates that this hiatus does not exist.

Faunas and age of the Warrior formation.—Correlation of the Warrior faunas with the standard Dresbachian faunizones is difficult; additional stratigraphic collecting and more detailed biostratigraphic study of existing collections are needed. Table 1 presents the vertical disposition of selected trilobites from collections of both Tasch and the writer. Species of *Coosella*, *Kingstonia*, *Blourtia*, and *Ankoura* are largely omitted because the genera mean little stratigraphically, and the writer's limited knowledge and material would make specific identification questionable. Tasch's published identifications have been relied upon heavily although the writer does not concur with all of them. For example, *Lonchocephalus waddlei* and *L. swartzi* Tasch are certainly congeneric if not conspecific with *Welleraspis jerseyensis* (Weller) (Howell, 1945, p. 2, pl. 1, figs. 1–4). Some

TABLE 1.—VERTICAL DISTRIBUTION OF WARRIOR TRILOBITES

NORTH Waddle Road (45-3w)	Warriorsmark	Williamsburg (47-24w)	Morrison Cove (47-1w) Schneider Farm	SOUTH Turnpike (47-11w)
0 Top of formation		(35) *Pemphigaspis bullata*		
		Lonchocephalus swartzi		
45–3w.2, Tasch 13.3		47–24w.3 *Kingstonia apion*		
(70) *Coosella* cf. *beltensis*		*Genevievella campbellina* (pyg.,		
Tricrepicephalus		Tasch, 1951, pl. 47, fig. 1)		
Crepicephalus (Tasch, pl. 46, fig. 12)		*Maryvillia* sp.		
100 *Maryvillia* sp. *Blountina*				(110) 47-11w.1
	(upper third, Loc. 24-f, ident. by Tasch)		(140) *Lonchocephalus swartzi*	*Lonchocephalus swartzi*
covered			47-1w.7 *Pemphigaspis bullata*	*Meteoraspis* sp.
	Genevievella campbellina		*Eshelmania* cf. *snoburgensis*	*Kingstonia apion*
	Bonneterrina aspinosa		(180) *Lonchocephalus swartzi*	*Maryvillia* sp.
200	*Ankoura* sp.		47-1w.6 *Pemphigaspis bullata*	
			(220) *Llanoaspidella warriorsmarkensis*	
			47-1w.5 *Lonchocephalus swartzi*	
			Pemphigaspis bullata	unexposed
			(275) *Lonchocephalus swartzi*	
covered			47-1w.1 *Genevievella campbellina*	
			Genevievella ? pennstatensis	
			(290) *Meteoraspis ?*	
300			*Genevievella ? pennstatensis*	
			Kingstonia sp.	
covered				
400				
Kingstonia apion (top occurrence, Tasch 12.97)				
Genevievella ? pennstatensis (top at 440, Tasch 12.66)			unexposed	unexposed
Genevievella campbellina (top at 450, Tasch 12.57)		*Lonchocephalus* sp.		
500				
700				
(720, Tasch, 11.12)	(lower Warrior, Loc. 24-g, ident. by		unexposed	unexposed
Genevievella campbellina	Tasch)			
800 *Kingstonia apion*	*Llanoaspidella warriorsmarkensis*			
900 (925, Old Quarry, 45-3w.0 and 0a)				
Pemphigaspis bullata (Ident. by Tasch)				
Lonchocephalus sp. undet. (none of Tasch's sp.)			unexposed	unexposed
Llanoaspidella warriorsmarkensis				
Millardia avitas, Kingstonia apion				
Genevievella campbellina				
925				
FEET	(basal Warrior faunas uncollected)			

doubt exists as to a distinction between *Welleraspis* and *Lonchocephalus; Welleraspis* evidently was incompletely studied by Kobayashi. Palmer (1951, p. 762, 763) has demonstrated that the cranidium *Hallaspis matutina* (Hall) is synonymous with *Pemphigaspis bullata* Hall and the writer has identified all his material as the latter.

Tasch (1951, p. 280–283) studied the Warrior fauna collected in a detailed measured section at Waddle Road junction (4503w) and attempted comparison with the type section along Warrior Creek, aided by several old collections made by Butts and others from this area. The old collections are only roughly placed stratigraphically, and subdivision of the formation into *Cedaria* and *Crepicephalus* zones of the standard Dresbachian section is not clear. Tasch presents evidence that the upper 70 feet of the Waddle Road section is of undoubted *Crepicephalus* age, and the writer concurs. A collection 35 feet below the top of the Warrior at Williamsburg (47-24w.3), about 20 miles south of the Waddle Road section, contains abundant specimens of *Blountina* and *Maryvillia* with *Pemphigaspis bullata* and is also considered of *Crepicephalus* zone age by the writer. At least the upper 100 feet of the Warrior formation should contain a *Crepicephalus* zone fauna everywhere along its outcrop.

The occurrence of *Eshelmania* cf. *E. snoburgensis* (also found in the *Pseudosaratogia magna* fauna of the basal Ore Hill member, Gatesburg formation) 140 feet below the top of the Warrior in Morrison Cove (47-1w.7) is noteworthy. This genus and *Cheilocephalus* are the only forms in the Appalachians which bridge the gap between Dresbachian and Franconian stages.

The middle and lower Warrior at Waddle is considered *Cedaria* zone equivalent by Tasch because species of *Ankoura*, *Genevievella*, and *Kingstonia walcotti* are abundant below his horizon 12.97. The present writer believes that a *Cedaria* zone age is not yet proved for any part of the Warrior formation. Tasch's species *Genevievella campbellina*, *G. ? pennstatensis*, *Llanoaspidella warriorsmarkensis*, and *Lonchocephalus swartzi*, present beneath his definite *Crepicephalus* beds, occur widely throughout the Warrior formation (Table 1), some from the lowest collection at the northernmost section (Waddle Road) to about 100 feet from the top southward in Morrison Cove where they are associated with *Pemphigaspis bullata*. The writer considers the oldest collection made in the Warrior to be his from the Old Quarry (45-3w.0 and 0a); the faunule may be of *Cedaria* zone age but is not definitely so. Preliminary identifications of this faunule by the writer were published by Tasch (1951, p. 306, addendum). Following further study and Tasch's recently published descriptions this is a complete list of the Old Quarry fossils in the writer's collection:

Present identification	*Former identification*
Blountia sp. (pygidium)	
Coosella sp.	
Genevievella campbellina Tasch	*Nixonella ?*
Kingstonia apion Walcott	
Llanoaspidella warriorsmarkensis Tasch	*Bolaspidella*
Lonchocephalus sp. undet. (different from higher species, *L. waddlei* and *L. swartzi*)	*Welleraspis*
Millardia avitas Walcott	
Semicirculurea sp. (gastropod)	

Tasch adds to this list a *Pemphigaspis* from the Pennsylvania State College collection, noting that this is an anomalous occurrence in a fauna with "a *Cedaria* zone cast". A. R. Palmer (U. S. Geological Survey, personal communication) after briefly studying specimens from this faunule believes on the basis of the species of *Coosella* and the presence of *Millardia avitas* that it represents either high *Cedaria* or low *Crepicephalus* zone or possibly a transition between the two. Many characteristic genera of the *Cedaria* zone are lacking in the fauna, and since the species of *Genevievella* and *Llanoaspidella* have been found so high in the southern sections, the writer considers an unquestioned *Cedaria* fauna yet to be proved for the Warrior formation. The lower 300 feet of the Warrior may well be of earliest Dresbachian and late Middle Cambrian age since the upper Pleasant Hill faunas belong to a zone low in the Albertan.

Gatesburg Formation

Nomenclature and location of exposures.—The Gatesburg formation was named from Gates-

burg Ridge in Center County, Pennsylvania by E. S. Moore (Butts, 1918, p. 587). The members are named by Butts in the same paper, and type localities indicated, but no lithologic descriptions are given. Further work in central

TABLE 2.—MEMBERS OF THE GATESBURG FORMATION

Larke dolomite-Stonehenge limestone	(Gasconade fauna of Canadian)	
Gatesburg formation 1700–1800 feet (Franconian and Trempealeauian Upper Cambrian)	Mines dolomite	250 feet
	Upper sandy Gatesburg member	650–700 feet
	Ore Hill member	200 feet
	Lower sandy Gatesburg member	400 feet
	Stacy dolomite	200–400 feet
Warrior formation (Dresbachian, Upper Cambrian).		

Pennsylvania resulted in general descriptions of the rock (Butts and Moore, 1936, p. 15–17; Butts, 1945, p. 2, 3; Butts *et al.*, 1939, p. 10–12), but the most detailed analysis of Gatesburg sediments has been made by Krynine (1941, 1946, 1948a).

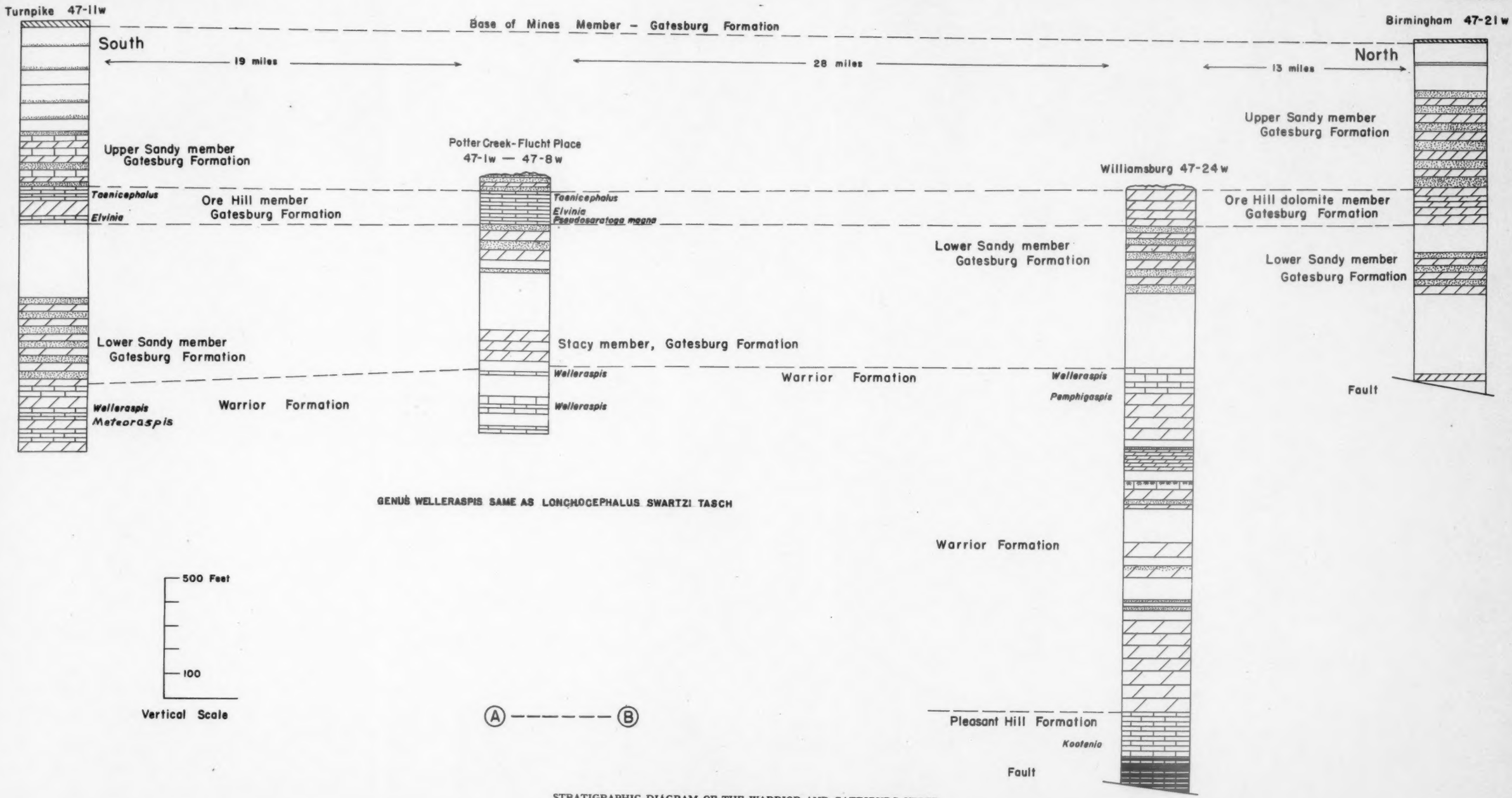
Butts divided the Gatesburg into four members and designated a thinner formation, the Mines dolomite, above them. Krynine and the writer include the Mines as a member of the Gatesburg (Table 2).

Only four sections of the Gatesburg formation are well enough exposed to warrant measuring within the Tyrone, Hollidaysburg, and Everett quadrangles (Pl. 1). By far the best exposed is the northernmost section located on the main line of the Pennsylvania Railroad on the Little Juniata River and in highway cuts just east of Birmingham Station (47-21w). The lower part of this section is cut out by reverse faulting, but the complete lower Gatesburg interval is present and fairly well exposed at the top of the long section at Williamsburg, 13 miles to the south (47-24w). Parts of the Stacy, lower sandy Gatesburg, and Ore Hill member are exposed at the south end of Morrison Cove, Everett quadrangle (Potter Creek-Flucht Place), and the southernmost section was measured in road cuts along the Pennsylvania Turnpike between Bedford and Everett stations. These sections trend southward for 62 miles through the three quadrangles.

Stacy member.—The lowest Gatesburg consists of massive dolomite lacking arenaceous layers. Butts (1945, p. 2) named these for exposures on Stacy knob a mile west of Larke road junction in the Huntingdon quadrangle and states that the member is nowhere well exposed. However, a fairly good idea of its content may be obtained from the road cut in the Everett quadrangle on Pennsylvania State Highway 868 where the road and Potter Creek leave Morrison Cove to enter the ridge of Gatesburg to the east. About 130 feet of alternating dark crystalline massive dolomite and less resistant thin-bedded lighter dolomite is exposed and is separated from the underlying Warrior limestone by 50 feet of covered strata.

Although Butts notes a thickness of 500 feet for the Stacy, the writer believes this excessive. The lowest orthoquartzite in the Gatesburg at the Williamsburg section (47-24w) occurs about 360 feet above the highest exposed Warrior. The two Everett quadrangle sections show only 100 or 200 feet of nonarenaceous dolomite at the base of the Gatesburg. This basal member is largely obliterated by faulting at Birmingham (47-21w).

Lower and Upper sandy members of the Gatesburg formation.—These members comprise most of the Gatesburg but are not given formal names. They are separated by the Ore Hill member. The two members are lithologically similar sequences of alternating orthoquartzites, pure and impure dolomites. The lower member is about 400 feet thick; the upper measures from 650 to 700 feet. Studies of the excellent exposures of the Birmingham section (47-21w) by Krynine and Pelto (Krynine, 1946, p. 2–7) have shown that several major rock types alternate cyclically within these members: (1) dark, thin-bedded, aphanitic silty dolomite weathering to a characteristic buff color; (2) medium to coarse-grained orthoquartzite; (3) thin-bedded finely crystalline and in part shaly dolomite; (4) massive black medium-crystalline dolomite; and (5) dolomitic cryptozoon stromatolite. At the Williamsburg section (47-24w) some red shale was found with the buff silty member of two cycles in the lower sandy Gatesburg member.



STRATIGRAPHIC DIAGRAM OF THE WARRIOR AND GATESBURG UPPER CAMBRIAN FORMATIONS

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Krynine has studied the Gatesburg orthoquartzite extensively. The schematic thin section (Krynine, 1948b, p. 152) is typical of this rock. The grains are moderately coarse to fine quartz sand, usually well rounded and frosted, and cemented by either dolomite or silica. There is little or no matrix. Much of this orthoquartzite is conglomeratic; the larger grains are small angular blocks of dolomite or fragments of other sedimentary rocks mixed with quartz grains of sand size. Beds are from 6 inches to 8 feet thick. Many are cross-bedded and ripple-marked.

The thin-bedded purer dolomite contains shale partings and is generally finer-grained than the highly characteristic more massive dolomite (fourth rock type) which is coarsely crystalline (2 mm average), oölitic, and has nodular chert. Many massive dolomite beds are 2–5 feet thick. The cryptozoon stromatolite, which seems confined to this lithotope, is considered a variety of it.

The cyclical arrangement of these rock types is easily discerned. There seems little difference in the type of cycle of the lower and upper sandy members. The following may be considered typical:

(1a) orthoquartzite, commonly conglomeratic at base *or* (1b), impure dolomite as below.

– – – – – – – – sedimentary break – – – – – – – –

(3) massive coarsely crystalline dolomite, commonly oölitic and cherty and with cryptozoon stromatolite at top; 3 to 9 feet thick.

(2) thin or medium-bedded dolomite, transition to above lithotope; several feet thick.

(1a) orthoquartzite as described above, conglomeratic at base; 2 to 5 feet thick.....*or* (1b) dark, thin-bedded, silty-shaly, buff-weathering aphanitic dolomite.

– – – – – – – – sedimentary break – – – – – – – –

(A complete cycle may range from about 3 to 25 feet thick.)

The two sandy Gatesburg members furnish abundant evidence of very shallow-water deposition. The blobs of dolomite and siltstone in the conglomeratic sandstones could have been transported only very short distances without disintegration. In some places this bottom scour and rapid reburial of loosened material formed flat pebble limestone or dolomite conglomerates in a carbonate instead of arenaceous matrix. At places oölite is intimately mixed with sand grains, and oölitic dolomite may be just as cross-bedded as the orthoquartzite. Oölite is considered to form in shallow, vigorously moving water with a high concentration of salts in solution (Hatch, Rastall, and Black, 1946, p. 178). A conjecture as to maximum depth of the water forming the dolomitic stromatolite is based on Cloud's statement (1942, p. 371) that "It is undoubtedly conservative to say that stromatolite-forming algae will not ordinarily flourish below a depth of about 30 meters in marine waters. . . ." The present consensus favors much shallower depths for abundance of blue-green (cryptozoon ?) algae so that water depositing the massive dolomite lithotope may have been commonly of tidal depth or at most only a few meters deep (Hatch, Rastall, and Black, 1938, p. 166; Hadding, 1933, p. 17, 18). Thus, the upper beds of the cycle, while undoubtedly clearer-water deposits than those lower, may not necessarily represent deeper-water deposition.

Sedimentary study of the orthoquartzites (Krynine, 1941, p. 1515) indicates a source area for the Gatesburg sediments that was previously subjected to severe chemical decomposition as well as mechanical attrition. Heavy-mineral residues show concentrates only of the highly resistant species, tourmaline and zircon. The paucity of finer debris within the formation may indicate that the source area furnished only a minimum of this material.

Ore Hill member.—The Ore Hill member is the only well-exposed part of the formation and because of its abundant fauna was more closely studied than other members. It was defined as a limestone by Butts (1918, p. 527). Publication of details concerning it dates from 1945 (Butts, 1945, p. 3), but field work was done in 1913 when Butts and Ulrich assembled a fairly representative collection of fossils from only two localities. The type section is in Ore Hill quarry about 300 yards northeast of a crossroads of Pennsylvania State Highway 686 and a road leading east to Ore Hill station Bloomfield township, Bedford County. The quarry has deteriorated, and the member is too poorly exposed to permit measuring of a complete section; physical stratigraphy does not

place the partial sequence exposed, but its fauna shows it to be the lower part of the member. A better and almost completely ex-

47-21w Birmingham, Tyrone quadrangle (complete section dolomite)

47-24w Williamsburg northern Hollidaysburg, quadrangle (complete section dolomite)

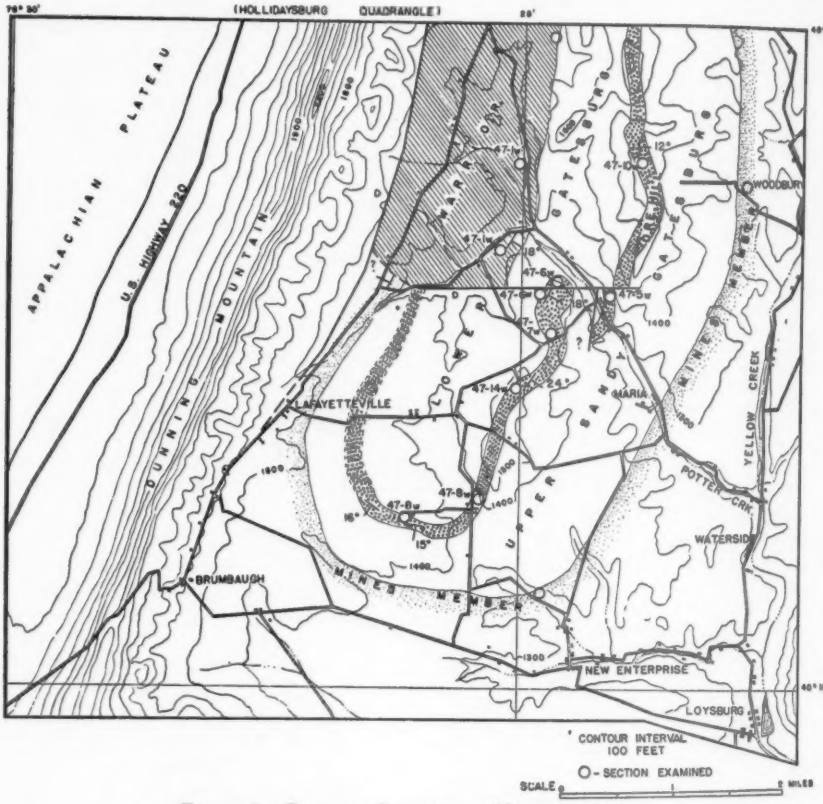

FIGURE 2.—CAMBRIAN OUTCROPS IN MORRISON COVE
Southern Nittany arch, Bedford County, NW ¼ Everett Quadrangle, Pennsylvania

posed section is Butts' and Ulrich's second fossil locality, the Drab section, 47-3w, U.S.N.M. loc. 107v, on slopes north of and in a low road cut south of an asphalt township road a quarter of a mile west of Drab-Beavertown, northern Huntingdon quadrangle, Pennsylvania.

The writer collected fossils from and measured the following sections of the Ore Hill member. (See also Wilson, 1951, p. 619–622; Figs. 2, 3, this paper.) Sections are listed proceeding southward.

47-3w Drab, northern Hollidaysburg quadrangle (complete section limestone)

45-2w Ore Hill quarry, southern Hollidaysburg quadrangle (partial section, lower Ore Hill)

47-10w West of Woodbury, top of ridge 2 miles west of village, northwest Everett quadrangle (partial section upper Ore Hill)

47-5w Potter Creek, northwest of Maria, 1½ miles, northwest Everett quadrangle (complete section, limestone)

47-6w South Potter Creek, southwest and across creek from 47-5w, northwest Everett

quadrangle (complete section, crossed by fault)

47-7w Cross roads, top of wooded ridge 1 mile northwest of Maria, above Potter Creek, southwest of it ½ mile, northwest Everett quadrangle (complete section, poorly exposed)

47-8w Flucht-Eschelmann quarries, two quarries ½ mile apart, on and south of county road, 2 miles north of New Interprise village, northwest Everett quadrangle (three partial sections connected stratigraphically, complete limestone sequence)

47-11w Turnpike (19 miles south of above section), cuts on south side of Pennsylvania Turnpike, 1 mile east of Midway Station, southwest Everett quadrangle (complete section poorly exposed, middle portion dolomite)

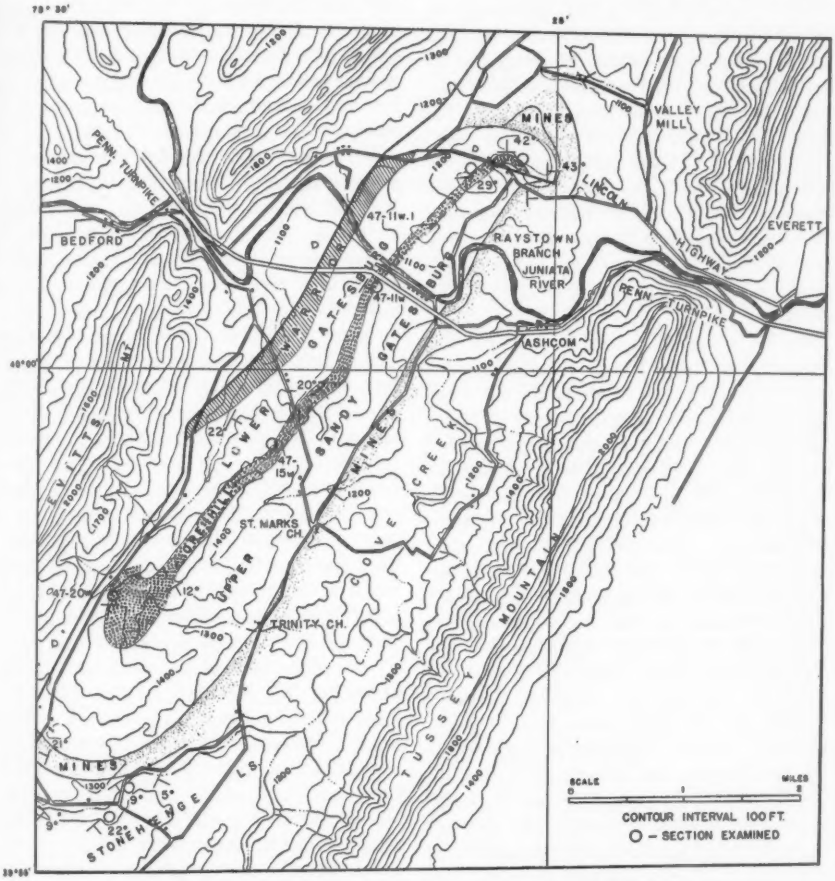

FIGURE 3.—CAMBRIAN OUTCROPS IN FRIENDS COVE, SNAKE SPRING VALLEY
Everett and Clearville quadrangles, Bedford County, Pennsylvania

47-14w Ruined house, in field top of ridge, 1 mile due west of Maria, northwest Everett quadrangle (partial section, upper Ore Hill)

47-15w Fickus quarry, 2 miles south of Hartley, ¼ mile west of road between Bedford and Rainesburg, northwest Clearville quadrangle (complete section, middle portion dolomite)

47-20w Imler quarry, east of county road, 2.5 miles southwest of Egolf Park, just west

of Evitt's Mountain, northwest Clearville quadrangle (complete section except for very base, middle portion dolomite) This section is 64 miles south of the northernmost section.

The writer redefines the Ore Hill to include the nonarenaceous dark crystalline dolomite of the Gatesburg found at the stratigraphic position of Butts' Ore Hill limestone from Williamsburg, Pennsylvania, northward through the Tyrone quadrangle and the Bellefonte quadrangle. The Ore Hill is easily mapped in the resistant limestone facies but may also be distinguished from the sandy Gatesburg members north of Williamsburg by the absence of any orthoquartzite throughout the dolomite sequence equivalent to the original Ore Hill limestone.

The 13 measured sections of the Ore Hill member show 7 types of carbonate rocks:

(1) Butts (1945, p. 3) describes the most common and characteristic lithofacies as "thin-bedded bluish to dark-gray, mostly fine-grained limestone with a few coarse layers." In a fresh quarry exposure the typical thin bedding is hardly noticeable, but on weathered surfaces the limestone is platy and flaggy, the individual layers 1 to 4 inches thick. Fifty cc of this limestone dissolved in HCl produced about 10 per cent, by volume, insoluble residue which was mostly black silt and clay with some very fine subangular sand. When fresh the limestone is black, but it weathers pale slate blue. In places dolomite and silt impurities produce yellow and white mottled weathered surfaces. Dolomitic portions weather out easily so that numerous beds are honeycombed or have deep thin gashes. This lithofacies is rarely fossiliferous. It occurs throughout the member centrally but is confined to the top and bottom of the unit at its southern extremity. It is the most resistant part of the Gatesburg formation.

(2) Some beds of dark medium-crystalline, massive limestone in the lowest part of the member are best exposed in quarries since they are not so resistant as the platy aphanitic limestone. The lithofacies bears the trilobites *Buttsia*, *Pseudosaratogia magna*, and *Eshelmania*, genera also found in the Conococheague limestone.

(3) At several localities (47-3w, 47-5w, 47-10w), the first-discussed lithofacies of the upper Ore Hill member bears very fine-laminated, minutely crystalline limestone. One sample shows about 80 laminae to the inch. Half are dolomite, lighter and much finer-grained than the dark limestone laminae with which they alternate. A thin section shows that the dark color of the limestone laminae is due to interstitial organic matter. The lighter layers are less resistant to weathering, and the natural surface of this rock appears "ribby."

(4) Coarsely crystalline fragmental limestone is also encountered in the Ore Hill. It is irregularly bedded, and the layers are highly variable in thickness but generally about an inch thick. These beds represent debris areas on the sea floor where trilobite and brachiopod fragments collected.

Two types of limestone intraformational conglomerate are present in the member. (5) The first, termed "pink pebble conglomerate" in the stratigraphic sections, is present in the first-described thin-bedded limestone in both top and bottom of the member, but is better developed and more characteristic in beds toward the base. Thin sections show the rock to be a calcirudite. It is generally fossiliferous, the trilobite parts occurring within the matrix which is gray, coarsely crystalline, and composed in large part of broken bits of trilobite tests and tiny pieces of the material that makes up the phenoclasts of the conglomerate. This lithofacies has furnished most of the fossil material so abundant in the member. The larger pebbles, from pin-point size up to several inches, are aphanitic, brick- or grayish-red pieces of limestone. They are of many shapes, predominantly flattish with the long axis and flat surfaces parallel to the bedding. Their rounded, contorted sides show that the limestone mud was malleable during burial. In one instance a trilobite test in the matrix left its impression on one of the pebbles showing that the piece was soft when deposited. The pinkish pebbles generally show a periphery of darker, more highly oxidized material even when the matrix is fresh limestone. There are no reddish limestones or dolomites within the Gatesburg formation except for these pebbles, and the writer considers their color the result of oxidation on mud flats just previous to a rapid burial. The mud flats must have been just barely above

water level and frequently washed over and moistened because none of the pebbles show any indication of being dried chippy material as in several other Cambrian intraformational conglomerates.

(6) The other type of limestone pebble conglomerate differs greatly from the first. It has smaller, black phenoclasts and is unfossiliferous. The larger grains are generally from a fourth of an inch down to pin-point size and are harder than the matrix which is dark, medium- to fine-crystalline limestone. The grains have a bumpy, nodular surface and are somewhat rounded.

(7) A massive medium-crystalline dolomite like that forming the top member of a typical sandy Gatesburg cycle is also present in the Ore Hill member at the southern extremity of its outcrop and completely replaces the limestone facies to the north. Most of this dolomite is dark, but some has a lighter hue.

The Ore Hill limestone-dolomite, like the sandy Gatesburg dolomite, contains much organic matter. When struck or crushed, the rock emits a strong odor of H_2S. Upon treatment with HCl, a black oily film appears and discolors the solution. The writer noted no cryptozoon stromatolite within the Ore Hill member, and, although sandy beds appear above and below it, none were noted within the unit.

The Ore Hill ranges from 130 to 213 feet thick, measured in six complete limestone sections through 55 miles of exposure. The dolomite sequences of the Ore Hill at the Williamsburg and Birmingham sections are about 160 feet thick.

Mines dolomite member.—The uppermost member of the Gatesburg formation consists of dark dolomite, siliceous oölite, and chert, much of it cryptozoon stromatolite. It lacks orthoquartzite. There is no reason for separating the Mines from the Gatesburg since two of its other members lack quartzite layers and some of its dolomites also contain chert like that of the Mines. Butts named the unit from incomplete exposures east of the Mines community in the Hollidaysburg quadrangle. The writer found but one good exposure of the member in the Everett quadrangle on a township road about 1½ miles northwest of New Enterprise toward Eschelmann quarry. Here the unusual beds of siliceous oölite with white matrix and black chert centers are present. Butts gives the thickness of the member as 250 feet, but the writer had no opportunity to check this. The residual cherts of the Mines member furnish an excellent stratigraphic marker and are highly useful in delineating the top of the Gatesburg along its outcrop.

Geographic expression of the Gatesburg formation.—Although more resistant than the limestone below and above it, the Gatesburg formation is very poorly exposed. The dolomite beds are particularly susceptible to chemical decomposition, and almost no natural exposures of them exist. As Butts has pointed out in several papers, the only residual weathering products are deep sandy soils containing numerous sandstone blocks. These litter the surface of the prominent densely wooded ridge produced by the formation. The crest of the ridge rises about 300 feet above the valley floors and is strikingly noticeable within the wide valleys or "coves" formed by the anticlines along the Nittany Arch in the Hollidaysburg, Everett, and Clearville quadrangles. It is supported by the Ore Hill member whose best exposures are generally found just below the highest points along it. The limestone is much more resistant than the dolomite beds in the sandy members below and above it. The Ore Hill has been quarried in past years for lime fertilizer, and numerous exposures of it occur in fresh and naturally weathered condition. The Mines dolomite weathers extensively, and the land is farmed in contrast to the wooded bramble-covered slope of the Gatesburg ridge.

Variation in the Gatesburg along strike.—Good exposures of the sandy members of the Gatesburg formation are few and incomplete, and beds well exposed at one locality are covered at another. Yet some evidence of variation in rock content may be seen between the two northern sections (Birmingham and Williamsburg) and the Turnpike exposures 50 miles to the south. The few measurable sections were plotted (Pl. 1) using the base of the Ore Hill as a key. Details of individual beds or cycles are not recognizable over distances greater than 10 miles in a north-south line. Purer, massive dolomite may become thin-bedded, and ortho-

quartzite beds may acquire intercalations of thin-bedded sandy buff dolomite, yet a general correlation of both arenaceous and pure dolomite units is possible.

TABLE 3.—PERCENTAGE OF ORTHOQUARTZITE BEDS IN GATESBURG SECTIONS

Section	North		South	
	Birmingham 47-21w	Williamsburg 47-24w	Eschelmann 47-8w	Turnpike 47-11w
Upper sandy member	17% (385 feet)		4% (62 feet)	10% (200 feet)
Lower sandy member	39% (150 feet)	18% (155 feet)	18% (140 feet)	10% (300 feet)

In the upper sandy Gatesburg of the Turnpike section, the 200 feet above the Ore Hill member includes numerous beds of limestone alternating with dolomite and orthoquartzite. In all, limestone constitutes 25 per cent of this part of the member. In the Flucht-Eschelmann area 19 miles northward, the lower 62 feet of upper sandy Gatesburg is exposed. These cyclical dolomites and orthoquartzites contain no limestone. Neither is there limestone in the Birmingham section 50 miles north of the Turnpike, but there is considerably more sand. Table 3 shows the percentage of the Gatesburg formation composed of orthoquartzite beds in parts of four sections. All beds exposed in the sections are considered; the sections are arranged geographically.

These data agree with studies by Krynine (1948a, p. 33) which show the Gatesburg to be much sandier toward the north and northwest.

The Ore Hill member shows lithofacies variation from north to south and is redefined to indicate this. In the vicinity of Larke, 3 miles south of Williamsburg (Butts, 1945, map), the limestone phase disappears northward. At this stratigraphic position northward, about 150 to 160 feet of massive dark crystalline dolomite is present. At Williamsburg the base of the Ore Hill dolomite is 635 feet above the base of the Gatesburg, at Birmingham the top of the member is 641 feet below the base of the Mines. These intervals may be checked satisfactorily against the complete Turnpike section. Two water wells 15 miles northeast of State College and just northeast of the Bellefonte quadrangle indicate that the Ore Hill dolomite interval may persist at least this far north; the top of the member is about 640 feet below the base of the Mines. This interval is based on correlation of the two wells which are 2200 feet apart in line of dip. P. D. Krynine kindly furnished the writer with logs of these wells, State College no. 2 and no. 5.

In the Drab section (47-3w) about 7 miles south of Williamsburg, the Ore Hill member is completely limestone, reasonably well exposed in slopes north of the road. In the excellent Potter Creek section (47-5w) of the northwest Everett quadrangle, an unexposed, nonresistant interval of 25 feet is present in the middle Ore Hill. This increases in the next section to the south (Eschelmann) and in the Turnpike and Fickus quarry sections 15 to 17 miles south of Potter Creek. Only the southernmost section (Imler quarry) completely exposes these beds (here 85 feet thick) and demonstrates that they constitute a southward-thickening unit of dolomite and light-brown limestone separating two more resistant platy dark aphanitic limestone sequences.

The lower of these thin-bedded dark limestones bears three or four beds of pink pebble conglomerates invariably containing fossils of the typical *Elvinia* fauna. These beds commonly overlie a dark, medium crystalline and massive limestone, widespread at the base of the member and containing the *Pseudosaratogia magna* fauna. The upper thin-bedded aphanitic limestone is present only from Potter Creek southward; 38 feet of more massive unfossiliferous dolomitic limestone appears at the top of the Ore Hill in the Drab section. Trilobites from beds immediately below this unit at Drab belong to the typical *Elvinia* fauna indicating that here almost the whole Ore Hill is of lower Franconian age. However, south of Potter Creek the higher unit of platy resistant limestone is persistent all the way south to the Imler quarry section and consistently bears trilobites of the *Conaspis* fauna and an occasional bed of pink pebble conglomerate. This upper unit is over 70 feet thick at Potter Creek but thins southward to only about 25 feet (Imler quarry) as the middle dolomite unit thickens (Pl. 2).

Stratigraphic boundaries.—The relation of the

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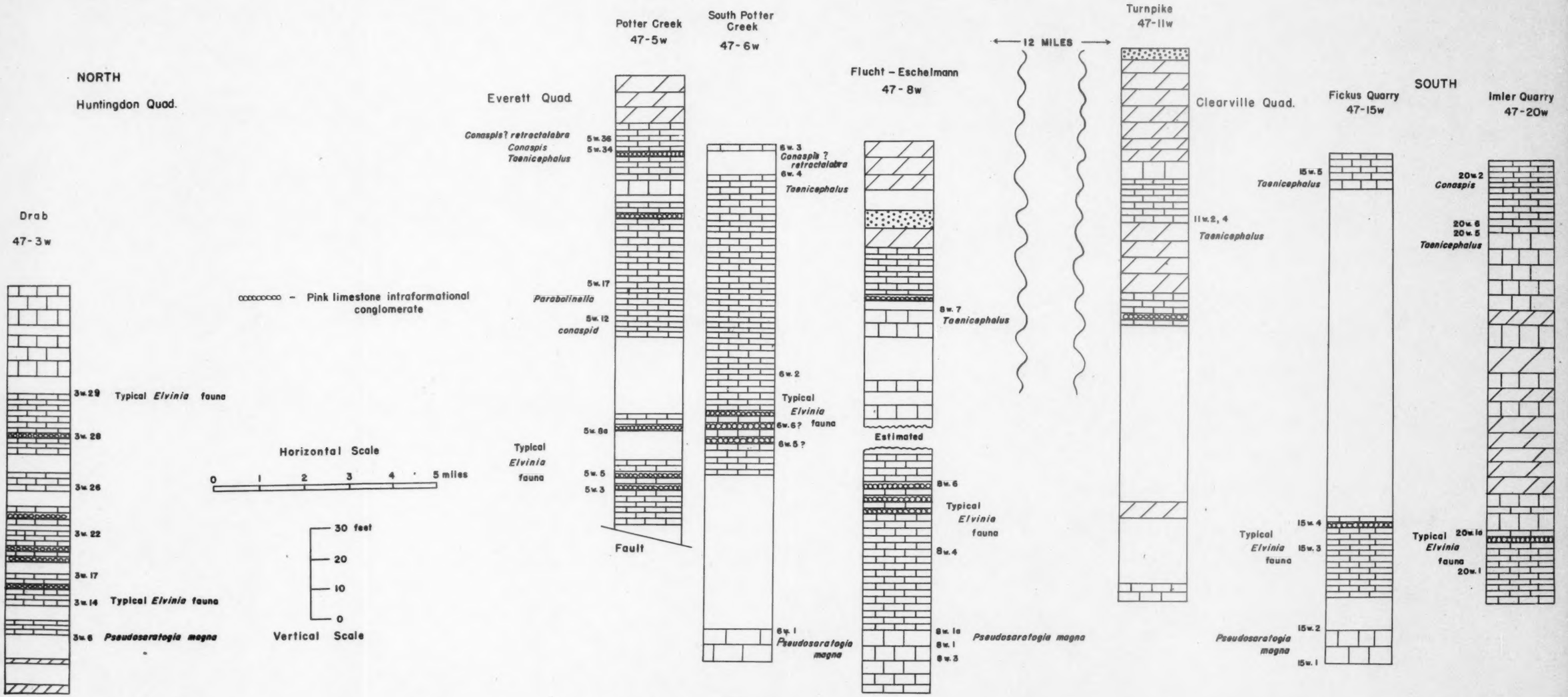


STRATIGRAPHIC SECTIONS OF THE ORE HILL MEMBER OF THE GATESBURG FORMATION

Gatesburg formation to overlying beds is obscure. The writer knows of no exposure of either the upper or lower contact of the Mines member of the Gatesburg. Yet the Mines is present completely along the north-south strike of the formation. It quite probably conformably overlies the upper sandy Gatesburg member and holds the same relation to the overlying Ordovician Larke dolomite or its limestone facies, the Stonehenge formation. This latter formation is present in the Bellefonte quadrangle but is replaced by the Larke dolomite southward through the Tyrone, Hollidaysburg, and Everett quadrangles. At the south end of the Gatesburg exposures, however, 4 miles north of Rainsburg, along Swamp Run branch of Cove Creek, southwest Clearville quadrangle, dark platy limestone containing orthoid brachiopods, nautiloids, and the trilobite *Symphysurina* overlies the Mines dolomite. This lower Canadian fauna helps identify the Stonehenge limestone at this southern locality.

Faunas and age of the Gatesburg formation.—The lower 650 feet of barren sandstone and dolomite (Stacy and lower sandy members) may represent at least in part the *Aphelaspis* faunizone of the standard Croixan sequence. The *Crepicephalus* fauna has been discovered as high as 60 feet from the top of the Warrior formation, but no genera of the *Aphelaspis* fauna have been collected at any locality within the central Appalachians.

Almost all fossils from the formation occur in the Ore Hill member: its limestone facies bears three Franconian trilobite faunas, along with both atremate and protremate brachiopods, gastropods, and several types of the pteropod-like genus, *Hyolithes*. Only the large and varied trilobite faunas have been studied (Wilson, 1951). They identify the *Pseudosaratogia magna*, *Elvinia*, and *Conaspis* faunizones in the central Appalachian area.

The most nearly complete development of the Ore Hill faunizones occurs in the Potter Creek sections, but the basal faunal unit found in the South Potter Creek section (47-6w.1) is separated by faulting from the well-known *Elvinia* and *Conaspis* zones so well represented north of Potter Creek (47-5w.3 and .5; 47-5w.17 through .36). The position of the two lower zones is checked in the Drab section (47-3w.6), and all three faunas are present in sequence in the Flucht-Eschelmann (47-8w.1, 8w.6, and 8w.7) and Fickus Quarry (47-15w.1, 15w.2, 15w.3, 15w.5) sections. The Imler Quarry section (47-20w.1 and 20w.5) shows the relation of the two upper zones only. These sections suggest that in the Drab section (47-3w) the upper beds containing the *Conaspis* faunizone are either thin and unfossiliferous or absent and that more sedimentation occurred there during the early Franconian since the over-all thickness of the member is normal at Drab. The *Elvinia* zone is persistent in the lower middle part of the Ore Hill associated with pink pebble conglomerates in platy aphanitic limestone. This zone grades downward into the lowest faunal unit.

1. *Pseudosaratogia magna* faunizone: This lowest faunal unit of the Ore Hill, about 10 to 15 feet thick, consists of dark, coarsely crystalline, massive limestone. It is marked by the teilzones of *Pseudosaratogia magna*, *Eshelmania snoburgensis*, and *Drabia curtoccipita*. The uppermost occurrence of these is overlapped by the lowest range and epibole of *Buttsia drabensis*. Some mid-continent *Elvinia* zone species are also present high in the unit (47-3w.6 and 47-15w.2): *Deadwoodia duris* (Walcott), *Dellea butlerensis* Frederickson, and rarely *Dellea suada* (Walcott). (*See also* 47-3w.6 in measured section.)

2. The typical *Elvinia* fauna of the mid-continent sections of Texas, Oklahoma, Missouri, and Wisconsin is present in the Ore Hill member with some modification and identifies the *Elvinia* faunizone. It consists of 27 trilobite species and 3 unassigned pygidia.

Species present throughout the mid-continent area:

Burnetia urania (Walcott)
Camaraspis convexa (Whitfield)
Camaraspoides berkeyi (Resser)
Cliffia lataegenae (Wilson)
Deadwoodia duris (Walcott)
Dellea suada (Walcott)
Dellea saratogoensis (Resser)
Dellea butlerensis Frederickson
Dokimocephalus intermedius (Resser)
Elvinia roemeri (Shumard)
Housia vacuna (Walcott)
Irvingella major Ulrich and Resser

* *Kindbladia wichitaensis* (Resser)
* *Plataspella anatina* (Resser)
Pterocephalia sanctisabae Roemer

Species limited to Pennsylvania, widespread *Elvinia* zone genera:

* *Bynumina terrenda* Wilson
Berkeia granulosa Wilson
Cheilocephalus buttsi Resser
Deckera completa Wilson
* *Kyphocephalus ponderosus* Wilson
* *Xenocheilos spineum* Wilson
Cheilocephalus sp. undet. (pygidium)

Genera and species seemingly confined to the Ore Hill:

Buttsia drabensis Wilson (also in Conococheague formation)
Drabia acroccipita Wilson
Drabia menusa Wilson
Pinctus latus Wilson
Pseudosaratogia lata Wilson
Pseudosaratogia bulbosa Wilson
Unassigned pygidia B, L, P

These species range between 10 and 100 feet above the base of the member, generally occurring in any one section through about 50 to 70 feet of flaggy dark aphanitic mottled limestone and commonly present in the matrix of the pink pebble conglomerates found in this part of the member. Each of these coquinas yields about a dozen species, with several forms abundant. Wave action has worn and broken the material and currents have probably sorted it.

The *Elvinia* fauna of the Ore Hill seems uniform both vertically and laterally with one exception. Collections from Drab at the northern end of the limestone facies outcrop (47-3w.17 through 3w.26) show species of *Plataspella*, *Kindbladia*, *Cheilocephalus*, and *Xenocheilos* through about 30 feet of beds 45 to 75 feet above the base of the member. With these and ranging lower but not higher are *Bynumina terrenda* and *Kyphocephalus ponderosus*. These six species are confined to the lower half of the *Elvinia* faunizone in the Ore Hill member and to the northern part of the outcrop belt. The species of *Plataspella*, *Kindbladia*, and *Cheilocephalus* were found only at Drab; the only other place where *Bynumina terrenda* and *Xenocheilos spineum* occur commonly is South Potter Creek (47-6w.5), and *Kyphocephalus ponderosus* only at Ore Hill quarry (45-2w).

These species are associated with other members of the *Elvinia* fauna, which occur both below and above the collection at Drab, but their presence together is interesting because they also occur together in mid-continent sections. Species of *Plataspella*, *Kindbladia*, *Cheilocephalus*, and *Xenocheilos* are common in Texas low in the *Elvinia* faunizone (Wilson, 1949, p. 30), and the writer has seen indication of the same association in sections from Oklahoma and Wyoming. The presence of this faunule in the Ore Hill section at Drab with and above genera which commonly overlie it in Texas could be attributed to migration and later occurrence of a favorable environment for these trilobites in Pennsylvania. In general these forms occupy the lower part of the *Elvinia* zone at Drab as they do in Texas. More sedimentation and a more extensive lower *Elvinia* zone development in the northern part of the Ore Hill limestone outcrop may be indicated; the faunule is unknown in the Potter Creek or other southern sections which expose the *Elvinia* zone well.

Very rare or absent in the Ore Hill member are certain lower Franconian genera which are abundant in many sections in the Cordilleran geosyncline: *Dokimocephalus*, *Iddingsia*, *Dunderbergia*, *Pterocephalia*, *Pterocephalina*, and *Parairvingella*. No accurate correlation of subdivisions of the *Elvinia* fauna between the geosynclines is possible because of lack of stratigraphic information on these beds in the West.

The *Irvingella major* faunizone (Wilson and Frederickson, 1950, p. 894) has not been located in the Ore Hill. No typical *Elvinia* fauna is found closer than 40 feet beneath the *Conaspis* fauna, and in most places a greater thickness of unfossiliferous dolomite intervenes, so that this secule of time may be represented by barren beds in the central Appalachian area.

3. The *Conaspis* faunizone: Platy aphanitic limestone bearing the uppermost faunal unit of the Ore Hill is present only south from Potter Creek where it reaches a maximum thickness of 78 feet. This unit thins southward to 24 feet, the lower part being replaced by dolomite and massive white limestone at Imler quarry (47-20w). The *Conaspis* faunizone has fewer fossil-

* Trilobite genera of the *Plataspella* faunule

iferous beds than the lower part of the Ore Hill member, and individual beds are less fossiliferous. There are only a few trilobite-bearing pink pebble limestone conglomerates. Collections from the upper Ore Hill beds commonly are from either the conglomerate or isolated patches in the dark platy limestone and consist of parts of a single species with rarely a cranidium of another form. Because of the few collections only a small number of species are described.

Only the Potter Creek (47-5w and 47-6w) and Imler quarry sections (47-20w) indicate a possible stratigraphic succession within the *Conaspis* faunizone. Other sections are incomplete. At Potter Creek, *Parabolinella occidentalis* Wilson with *Wilbernia* occurs lowest followed by *Taenicephalus shumardi* (Hall) with *Wilbernia*. Still higher *Conaspis retractalabra* Wilson and *Conaspis perseus* (Hall) are added to the *Taenicephalus fauna* (47-5w.34, 47-.5w.36, 47-5w.3). At Imler quarry (47-20w.2, 47-20w.5–6) the *Parabolinella* fauna is absent probably because the dark platy aphanitic limestone lithofacies of the lower part of the faunizone is replaced here by massive unfossiliferous limestone and dolomite. Here the lowest fossiliferous beds bear *Taenicephalus shumardi* and *Pelagiella* sp., and the highest fauna consists of *Conaspis perseus* alone. Different species of *Taenicephalus* and *Conaspis* occur elsewhere in incomplete sections.

As in mid-continent sections, the lowest occurrence of abundant early orthoid brachiopods coincides with the appearance of the *Conaspis* fauna. Dr. G. A. Cooper examined the best material and tentatively referred some of the specimens to *Eoorthis* and *Apheorthis*, but the material is too poor for certain identification.

The Ore Hill faunas are lower and Middle Franconian. Butts has implied that Trempealeauian trilobites known in the Hoyt formation of New York are also present with the *Elvinia* zone faunas (Butts *et al.*, 1939, p. 12; Butts, 1945, p. 3). These statements were based on early identifications by Ulrich of the trilobites *Saratogia* and *Plethopeltis* and the gastropod *Pelagiella* hoyti (Walcott). *Saratogia* resembles the Ore Hill species *Pseudosaratogia magna* Wilson, but the differences have been discussed (Wilson, 1951, p. 648). The writer has been unable to find *Plethopeltis* in the Butts and Ulrich collection or in his own material. *Pelagiella* proves little since species of the genus are present generally in the mid-continent Franconian.

The only Trempealeauian fauna now known from the Gatesburg occurs in the Mines dolomite member. Butts, who defined the Mines as a separate formation, considered it of the same age as the Van Buren dolomite (lowest Gasconade of Missouri (1945, p. 3) on the basis of his discovery of the gastropods *Sinuopea vera* and *S. planibasalis* (Ulrich) just south of Williamsburg, Pennsylvania. However, since the writer has discovered *Sinuopea* cf. *vera* with a species of *Prosaukia* in the Conococheague limestone of the Kendall Farm section at McConnellsburg Cove, this species trangresses the systemic boundary, and the Mines may well be Cambrian.

Sections of Warrior and Gatesburg Formations

LOCALITY 45-3w, Waddle Road Section of Warrior Formation: Along U. S. Route 322, 5 miles west of State College, Pennsylvania. From the Waddle road junction east along the highway past Scotia road junction and up the hill to an obscure contact with the Gatesburg. There are three cuts along the highway. The upper part of the section is in the easternmost, highest cut. A detailed measured section through these cuts published by Tasch (1951, p. 276–278) agrees rather well with the writer's. However, since the writer's section begins lower than Tasch's (at the Old Quarry locality) and one interval differs appreciably, an abbreviated section is given here.

LOCALITY 47-21w, Birmingham Station east of Tyrone, Pennsylvania: Along the Little Juniata River on the main line of the Pennsylvania Railroad about 4 miles east of Tyrone and just east of Birmingham Station in bluffs and road cuts along the highway from Tyrone to Union Furnace. This is the most nearly complete section although little of the Stacy member is exposed and there are numerous gaps. The section was first mentioned and photographed by Butts (Butts *et al.*, 1939, p. 5) and was studied by C. R. Pelto under direction of

TOP OF THE WARRIOR FORMATION: PLACED SOMEWHAT INACCURATELY BELOW BASE OF SANDY DOLOMITE FLOAT ASSIGNED TO THE GATESBURG

Description of interval	Tasch's bed number	Feet
Mostly covered, some platy limestone		73.0
Platy limestone		10.0
45-3w.2, *Coosella, Blountia, Maryvillia, Tricrepicephalus, Blountina*	Within 13.1	
Massive limestone		3.0
Platy limestone and dolomite, some *Cryptozoon undulatum* type stromatolite	Within 13.1	4.5
Dolomite	Base of 13.1	1.0
Total Warrior from top to base of highest road cut (Tasch's total is 107.3 feet) (Dip is 30°, S. 30° E.)		91.5
Interval covered except for a 20-foot ledge in field about 15 feet from top (Dip at base is 33½°, S. 25° E.) Tasch's total for this interval is 287.5 feet	13.0	221.0
Alternating massive and platy limestone and dolomite. Described in detail by Tasch. 45-3w.1 (Tasch's 12.49) containing *Kingstonia, Coosella* (?), and *Maryvillia* 44.5 feet above base and 10 feet below an algal bed with a white "W" painted on it. Tasch's total for interval is 104.7 feet (Base of middle road cut)	12.102 to 12.1	122.0
Covered interval, dip at top of lower cut is 17°, S. 25° E. and at base of middle cut 35° S. 40° E.; consequently, an average dip of 26° was used to calculate thickness. As in the lower covered interval ground distance in line of dip was taped. (Tasch's total for this interval, 176 feet)	12.0 to 11.41	172.0
Top of lower road cut		
Alternating beds of massive and platy dolomite and limestone, with some dolomitic shale and siltstone. (Tasch's total to base of lower road cut, 94 feet.)	11.40 to 11.1 Base of Tasch's section	105.0
Covered interval, error in meeting base of lower cut about 5 feet		118.5
Covered interval except top, which is dark, resistant aphanitic dolomite, measured up a steep slope above the Old Quarry		88.0
Dark crystalline limestone, exposed in Old Quarry, 100 yards north of the Waddle road junction with highway. 47-3w.0, *Kingstonia apion, Coosella, Blountia, Genevievella campbellina*		7.0
47-3w.0a Float at about same horizon in side road just north of highway. Beds exposed in this road bed show dips consistent with those farther east, and no evidence of reverse faulting was noted. *Blountia* sp, *Millardia avitas, Llanoaspidella warriorsmarkensis, Coosella* sp. *Lonchocephalus* sp. undet., *Kingstonia apion*		
Total Warrior formation measured		925.0

Although Butts and Moore (1936, map) show a fold axis in the base of the section the dips below the lower cut are consistent (15°–17° S. 22½° E.), and the fold axis is probably west of the section.

P. D. Krynine in 1942. The Gatesburg formation here has been thrust over Silurian and Upper Ordovician formations, and the plane of thrust cuts through beds of the Stacy member exposed in railroad cuts across the river and below the highway. The higher Gatesburg with well-developed cyclic deposition is seen in three major cuts along the highway.

Upper Sandy Member of the Gatesburg Formation: Top of the Gatesburg Taken as Top Exposed Orthoquartzite	Feet
Covered interval with a few weathered quartzites exposed	186
82. * Mostly weathered quartzite	15
81. Covered interval	4
80. Massive dolomite	7
79. Quartzite	2
78. Massive dolomite	9
77. Quartzite	1.5
76. Oölitic dolomite	1.5
75. Quartzite with dolomite pebble conglomerate	1
74. Dolomite	1
73. Coarsely oölitic dolomite	1
72. Buff silty dolomite	.5
71. Cryptozoon	.5
70. Dark oölitic dolomite	3
69. Buff silty laminated dolomite	1
68. Oölitic dolomite	3.5
67. Quartzite	3
66. Buff silty dolomite with cryptozoon	1.5
65. Oölitic dolomite	1
64. Buff silty dolomite	.5
63. Buff silty dolomite with cryptozoon	1.5
62. Oölitic dolomite	1.5
61. Quartzite and dolomite pebble conglomerate	2
60. Massive oölitic dark crystalline dolomite	12
Total Upper Sandy Gatesburg above cycles of Krynine and Pelto	247.5

A crushed fault zone; it is believed that only a little of the section is missing or repeated here because

(1) Zone of crushing is only 7 feet wide.
(2) Angle of thrust and dip of beds is same.
(3) Beds above and below show no repetition.

Cycle 53 of Krynine and Pelto (1942)	
Massive dark dolomite	5
Quartzite	2
Buff silty dolomite	9
Cycle 52	
Dark massive dolomite	6
Quartzite	2
Buff silty dolomite, arenaceous	1
Cycle 51	
Massive dark dolomite	4.5
Quartzite with silty layers, conglomeratic	2.5

* Imperfectly developed cycles.

	Feet
Cycle 50	
Dark massive dolomite	7
Quartzite	2.5
Buff silty dolomite, quartzite, conglomeratic	2
Cycle 49	
Thin-bedded dark dolomite	2
Quartzite, somewhat silty with conglomerate	1.5
Cycle 48	
* Dark massive dolomite	4
Buff silty dolomite	7
Cycle 47	
* Dark massive dolomite	4
Buff silty dolomite	4.5
Cycle 46	
Dark massive dolomite	2
Conglomeratic quartzite	3
Cycle 45	
* Dark massive dolomite, thin-bedded at top	2.5
Buff silty dolomite	3.5
Cycle 44	
Dark massive dolomite, cherty and oölitic	1.5
Conglomeratic quartzite	1.5
Cycle 43	
Dark massive dolomite	1
Conglomeratic quartzite silty at base	3
Cycle 42	
Dark massive oölitic dolomite	2
Conglomeratic quartzite	1.5
Buff silty dolomite	5
Cycle 41	
* Massive dark dolomite, conglomeratic at base	2
Buff silty dolomite	1.5
Cycle 40	
* Oölitic dark massive dolomite	3
Buff silty dolomite	2
Cycle 39	
Dark massive dolomite	3
Quartzite	5
Buff silty dolomite	3
Cycle 38	
Dark massive dolomite	3
Conglomeratic quartzite	1
Buff silty dolomite, oölitic at base	4
Cycle 37	
Dark massive dolomite, oölitic at top	3
Quartzite, conglomeratic and silty at base	1
Cycle 36	
* Dark massive dolomite	5
Buff silty dolomite	2.5
Cycle 35	
* Thin-bedded dark dolomite	7
Dark massive dolomite	5
Thin-bedded dark dolomite	3.5
Buff silty and thin-bedded dark dolomite	1

	Feet
Cycle 34	
* Dark massive dolomite	2
Thin-bedded dark dolomite, silty, conglomeratic and oölitic at base	1.5
Cycle 33	
Dark massive dolomite, arenaceous at base	3
Buff silty dolomite	9.5
Cycle 32	
Dark massive dolomite	6
Quartzite	9
Buff silty dolomite	1.5
Cycle 31	
* Thin-bedded dark dolomite, at base silty and oölitic	6
Cycle 30	
Thin-bedded dark dolomite with cryptozoon, oölitic at base	2.5
Silty, conglomeratic quartzite	.5
Total Upper Sandy member of Gatesburg in Cycles of Krynine and Pelto	189.0

The above section ends at the base of the easternmost road cut exposing the Gatesburg. The following section is a continuation measured above the road on the wooded slopes north of the highway. It is less well exposed and difficult to measure. The degree of accuracy is less than that of Krynine and Pelto (1942).

59. Massive cryptozoon dolomite	2.5
58. Massive cross-bedded and laminated dolomite	4
57. Quartzite with pebble conglomerate, thin-bedded	5
56. Laminated, thin-bedded finely crystalline dolomite	6.5
55. Massive dolomite	2
54. Cross-bedded quartzite	4
53. Dolomitic siltstone, thin laminae, buff silty dolomite	1.5
52. Massive dolomite	1
51. Quartzite	2.5
50. Aphanitic, light-colored very thin-bedded buff silty dolomite	2
49. *Massive coarsely crystalline dolomite	5
48. Crystalline dolomite with cryptozoon	4.5
47. Quartzite	1.5
46. Buff silty laminated finely crystalline dolomite	1.5
45. Quartzite	3
44. Massive crystalline dolomite	2
43. Covered interval	18
42. Quartzite	5
41. Light-colored cryptozoon dolomite	3
40. Dark dolomite, mostly covered	5.5
39. Massive and laminated dolomite	12
38. Quartzite	3
37. Finely laminated, platy-weathering silty dolomite	2
36. Covered interval	9.5
35. Massive dark crystalline dolomite	2
34. Intraformational dolomite pebble conglomerate	1.5
33. Quartzite	2
32. Platy thin-bedded buff silty dolomite	3
31. Covered interval	22
30. Thin-bedded dark dolomite	9
29. Laminated finely crystalline dolomite	5
28. Coarsely crystalline dolomite	2
27. Quartzite with dolomite pebble conglomerate in base	2
26. Fine grain, light-colored, thin-bedded dolomite	1
25. Finely crystalline dolomite, chert stringers in bottom third	10
24. Massive dolomite capped by cryptozoon and oölite	3.5
23. Soft weathering covered interval	2
22. Massive cross-bedded quartzite	3
21. Covered interval	5
20. Buff silty dolomite	4
19. Covered interval	5
18. Quartzite (arenaceous dolomite at top)	4
17. Laminated dark finely crystalline dolomite	4
16. Covered interval	3
15. Massive finely crystalline dolomite	1
14. Soft-weathering purple platy dolomite	3.5
13. Dolomitic quartzite	.5
Upper Sandy Gatesburg member below Cycle 30	204.5
Total Upper Sandy Gatesburg member	641.0

Ore Hill member of the Gatesburg formation: This with the sandy Gatesburg beds immediately above exposed in the middle road cut of the highway.

12. Dark finely crystalline dolomite; laminated in undulation; some edgewise conglomerate, with pink pebbles of dolomite	7.5
11. Covered, soft-weathering beds	3
10. Finely crystalline dolomite with silty undulating laminae	9.5
9. Finely crystalline gray dolomite in beds of 2 to 3 inches, undulating laminae. Looks like typical Ore Hill upper beds which are dolomitized	1.5
Upper Ore Hill dolomite	21.5
8. Massive dark crystalline dolomite	54.5
7. Covered interval, soft-weathering	3

	Feet
6. Coarsely crystalline, light-gray dolomite with linguloid brachipods in top	28.5
5. Covered interval	3.5
4. Massive coarsely crystalline dolomite	8
3. Covered interval	2.5
2. Dark crystalline dolomite	1.5
Dip here about 21° E., above beds measured perpendicular to bedding from side. Measurement is not considered as accurate as the other parts of the section.	
1. This interval only approximately right, dark coarsely crystalline dolomite. Basal bed weathers out into huge squarish blocks with algal heads in checkerboard pattern on their upper surface	40.0
Total coarsely crystalline Ore Hill dolomite	141.5
Total Ore Hill dolomite	163
Lower Sandy Member of the Gatesburg formation:	
Covered interval (approximately)	120
Massive buff silty dolomite	2
Quartzite	1
Medium to thin-bedded dolomite	4
Quartzite	1
Shaly thin-bedded buff silty dolomite	2
Quartzite	4
Buff silty medium-bedded dolomite	3
Quartzite	2
Buff silty dolomite and covered	3
Medium-bedded quartzite	4
Buff silty, platy, shaly dolomite	2.5
A small reverse fault with a displacement of 3 feet. The last few beds above and all that follows are exposed in the westernmost of the three road cuts.	
Buff silty medium-bedded dolomite	3
Total Lower Sandy member of the Gatesburg above cycles of Krynine and Pelto	151.5
Cycle 29 of Krynine and Pelto (1942)	
Massive dark dolomite	3
Quartzite	2
Buff silty dolomite	6
Cycle 28	
* Quartzite with siltstone, buff silty dolomite at base	3
Cycle 27	
* Quartzite, silty at base	3
Cycle 26	
Very thin-bedded black dolomite, appearing like black shale	6
Quartzite, conglomeratic in places	6
Cycle 25	
Cherty thin-bedded dark dolomite	1
Cherty, conglomeratic silty quartzite	1
Cycle 24	
Thin-bedded dark dolomite	2
Conglomeratic quartzite	3
Cherty, conglomeratic silty quartzite	1.5
Buff silty thin-bedded dolomite	1.5
Cycle 23	
Massive dark dolomite, oölitic at top	3
Conglomeratic silty quartzite	1
Cycle 22	
Thin-bedded dark dolomite	1.5
Conglomeratic cherty and oölitic quartzite	1.5
Cycle 21	
* Massive dark dolomite, sandy and cherty at base	3
Cycle 20	
* Massive dark dolomite, sandy and cherty at base	3
Cycle 19	
Dark massive dolomite	2.5
Quartzite	2
Cycle 18	
Thin-bedded and massive dark dolomite	2.5
Quartzite	.5
Cycle 17	
Dark massive dolomite	2
Cherty and conglomeratic quartzite	1
Cycle 16	
Dark massive dolomite	1.5
Quartzite, at base oölitic and conglomeratic and silty	8
Cycle 15	
Massive dark cherty dolomite, oölitic at base	2
Cherty conglomeratic quartzite, silty at base	4.5
Cycle 14	
* Dark, thin-bedded dolomite, silty at base	12
Cycle 13	
Dark massive cherty dolomite	3
Quartzite	3
Cycle 12	
* Dark massive dolomite, sandy at base	3.5
Cycle 11	
Dark, massive cherty dolomite	2.5
Quartzite, silty at base	2
Cycle 10	
Massive cherty dolomite	2.5
Silty quartzite	1
Cycle 9	
Dark massive dolomite with cryptozoon at top. (This is the lowest bed exposed at the east end of the railroad bridge across Little Juniata River)	6.0
Quartzite, silty at base	6.5
Cycle 8	
Dark massive dolomite	1.0
Conglomeratic quartzite, silty at base	2.5

	Feet
Cycle 7	
Dark massive dolomite	2.0
Conglomeratic quartzite, cherty and silty at base	2.0
Cycle 6	
Dark massive dolomite, cherty at top	2.5
Conglomeratic sandy dolomite	1.5
Cycle 5	
* Massive dark dolomite, cryptozoon at top and base, some conglomeratic sandy dolomite	5.0
This is base of westernmost road cut on the highway.	
Total Lower Sandy member of the Gatesburg formation measured by Krynine and Pelto	137.0
Cycle 9 may be traced to the base of the railroad cut from the cut on the highway. From this point an estimate of covered beds to the lowest Gatesburg may be made. Excluding Cycles 8-5 exposed on highway this amounts to	284.0
Stacy dolomite member of the Gatesburg formation:	
Cycle 4 of Krynine and Pelto (1942)	
Dark massive dolomite, cherty at top	2.5
Dark thin-bedded dolomite, silty at base	9.0
Cycle 3	
Dark thin-bedded dolomite, silty at base, massive at top	7.0
Cycle 2	
Dark massive oölitic and cherty dolomite	1.5
Dark thin-bedded dolomite, silty at base	6.0
Cycle 1	
Dark massive dolomite, cherty and cryptozoon at top	2.5
Thin-bedded dark dolomite	4.0
Total member Stacy measured by Krynine and Pelto	32.5
Considerably more Stacy dolomite is exposed in the railroad cut, but it is close to the thrust plane and too deformed for accurate measurement.	
Total thicknesses of Gatesburg members (exclusive of Mines)	
Upper sandy member	641
Ore Hill dolomite	163
Lower sandy member	288
Covered interval, lower part probably Stacy member	284
Stacy dolomite (incomplete)	32
Total Gatesburg exposed below Mines member	1408.0

LOCALITY 47-24w, Williamsburg: Along crest of the bluffs on north side of Frankstown branch of Juniata River from the faulted contact of the Pleasant Hill limestone and Canadian dolomites about 1 mile west of Williamsburg to within outskirts of the town. The section begins in the Pleasant Hill formation above the river bend where the fault crosses. A small draw follows the fault up from the river on the north bank; another runs up the bluff at the base of the Warrior formation. The Warrior-Gatesburg contact about a quarter of a mile east is obscured by a large covered interval where a power line crosses the bluffs but lies east of the high dolomite bluff above the dam. The striking "ribs" up the face of the bluff are beds in the Warrior. The lower sandy Gatesburg occurs just across the river from the sewerage outlet of a factory, and the cycles in this member are exposed in the bluff here. The crest of this ridge is supported by massive dolomites of the Ore Hill member. The upper Gatesburg is poorly exposed, and its upper contact is not clear. The section was roughly measured by Butts (1945, p. 2) and given as the best exposure of the Warrior formation. The writer remeasured this section and adds about 100 feet to the Warrior formation. Distance was chained, dips measured with the Brunton.

ORE HILL DOLOMITE MEMBER OF THE GATESBURG FORMATION

	Feet
(Crest of bluff just north of sewerage outlet)	
Massive dolomite	22.0
Sandy dolomite	1.0
Covered interval	17.0
Dark massive dolomite	42.0
Massive dark crystalline dolomite	17.0
Covered interval	10.0
Dark crystalline massive dolomite, in part oölitic	42.0
Total Ore Hill dolomite member	151.0

LOWER SANDY MEMBER OF THE GATESBURG FORMATION

The following 278 feet is so clastic that the Gatesburg cycles as developed at the Birmingham section are hardly more than an alternation of quartzite and silty dolomite beds. Comparison of the stratigraphic interval in the two sections based upon the base of the Ore Hill member shows that only about 30 feet of the well-exposed lower Gatesburg at Williamsburg overlaps with the well-exposed part of the formation at Birmingham, from about cycle 28 up through the beds added by the writer above cycle 29. The rock of this interval is closely similar in both sec-

tions; there is a minimum development of massive dolomite.

Buff silty dolomite leading up to sandstone	2.0
Thin shaly dolomite	5.5
Dark massive, thin-bedded dolomite	7.5
Buff silty dolomite	1.0
Dark dolomite	3.0
Buff silty dolomite	2.5
Massive dolomite	6.5
Platy dolomite	5.5
Buff shaly dolomite	9.0
Quartzite	1.0
Buff shaly dolomite	3.0
Quartzite	3.0
Buff silty dolomite	2.0
Quartzite	2.0
Buff massive dolomite	2.5
Buff shaly dolomite with red shale	6.0
Cherty oölitic dolomite	1.0
Quartzite	1.0
Massive buff-colored dolomite	11.0
Buff shaly dolomite with a little red shale	3.5
Buff massive dolomite	1.5
Buff shaly dolomite	1.0
Quartzite	1.0
Buff shaly dolomite	2.0
Bright-red shale	1.0
Buff shaly dolomite	4.0
Dark sandy dolomite	.5
Soft silty dolomite	.5
Quartzite	3.0
Thin-bedded soft silty dolomite	3.0
Conglomeratic quartzite	2.0
Soft dolomite	6.0
Fine dolomitic quartzite	1.0
Dark oölitic dolomite	7.5
Conglomeratic quartzite	1.0
Dark finely crystalline dolomite	.5
Covered interval	9.0
Cross-bedded conglomeratic quartzite	5.0
Silty soft thin-bedded dolomite	4.0
Quartzite, some cross-bedded and conglomeratic	4.0
Thin-bedded silty dolomite	7.0
Dark finely oölitic dolomite	6.0
Soft shaly-silty dolomite	2.5
Dark finely crystalline dolomite, massive	3.0
Quartzite grading up to sandy dolomite, sandy oölitic dolomite	5.0
Total well exposed lower sandy Gatesburg	159.5
Covered interval along river bank	73.0
Weathered quartzite bed followed down to river from crest of bluff, possible 10–15 feet error here	
Covered interval	44.0
Weathered quartzite bed	2.0
Covered interval where power line crosses	356.0
Total unexposed lower Gatesburg	475.0
Total lower Sandy Gatesburg and Stacy members	634.5

Warrior Formation

Platy aphanitic buff silty dolomite	10.0
Dolomite (dip 57°)	1.0
Thin-bedded limestone weathering platy and nodular	22.0
47-24w.2, *Lonchocephalus, Pemphigaspis, Coosella, Kingstonia, Blountia*	
Thin-bedded limestone	20.0
Covered interval (Possible top of Butts' Section)	17.0
Massive dolomite	29.5
Massive dolomite, forms bluff over dam, cryptozoon at the top, (dip 52½°)	161.0
Covered interval	28.5
Thin sandstone capped by cryptozoon	1.5
Platy aphanitic dolomite (dip 54°)	10.0
Platy aphanitic dolomite, light-weathering, medium-bedded	97.0
Covered interval	34.0
Dolomitic limestone with cryptozoon	2.5
Platy buff silty dolomite	2.0
Cryptozoon limestone (dip 57½°)	1.0
Dolomite	3.0
White-weathering fine-grained dolomite	40.0
Platy dolomite	1.0
Covered but float shows dolomite	19.0
Oölitic slabby limestone and dolomite	10.0
47-24w.1, *Lonchocephalus swartzi*, float, placed here	
Sandstone	1.0
Pure platy dolomite (dip 48½°)	17.0
Covered interval	144.0
Massive dark dolomite	25.0
Massive dark crystalline dolomite	32.5
Covered interval (dip changes to vertical here from 43°)	30.0
Sandy dolomite	3.0
Vertical sandstone with dolomite matrix, Eastern Indian Rock	4.0
Sandstone with some dolomite	31.0
Covered interval	81.5
Sandstone ridge, Western Indian Rock	5.5
Covered interval	16.0
Thin-bedded sandstone	2.0
Massive dark dolomite, some platy, light- and dark-colored dolomite	27.0
Dolomite	18.5
Covered interval	77.0
Resistant dolomite	20.0
Covered interval (dip vertical)	75.0
White dolomite and dark crystalline dolomite	39.0
Massive white-weathering light-colored dolomite, nicely crystalline, medium-grained	50.0
Covered interval	60.0
White-weathering dolomite	73.0
Total Warrior formation	1342.0

PLEASANT HILL FORMATION

Sandy shale and dolomitic limestone	20.0
Limestone, weathering into honey-comb, worm-eaten surfaces (Gully)	54.0
Platy impure limestone	16.0
47-24w.0, *Kootenia*	
Platy to medium-bedded limestone, less silty than below, weathering to worm-eaten beds	84
Sandy shale and dolomitic limestone (Dip 82°)	15
Upper member of Pleasant Hill	189.0
Calcareous or dolomitic shales, brown weathering, silty, and platy or fissile, shaly limestone toward the top (Dip 60°)	158
Total Pleasant Hill exposed (Fault cuts out base of formation)	347.0

LOCALITY 47-11w, Pennsylvania Turnpike: In road cuts along the Pennsylvania Turnpike about 1 mile east of the Midway Station and Restaurant. The section begins just west of the first wooded ridge east of Bedford Gap. Here the Warrior formation is exposed in road ditches and in a low cut on the highway. The section continues along the Turnpike for about 1 mile east. The distance was chained; the beds dip S. 30° E. to S. 60° E. at from 15° to 25°. The Turnpike trends first S. 77° E. and turns to about S. 25° E. Dips steepen eastward. The lower part of the Gatesburg is weathered in the westernmost cut, but the upper part is well-exposed for several hundred feet above the Ore Hill member. It is not known whether the complete upper sandy member of the Gatesburg is represented, but a thickness estimate was made up through the eastern edge of the wooded ridge. (The section in the meadows beyond is too badly covered for description.) This section shows the presence of limestone in the upper sandy member of the Gatesburg and a diminution of the orthoquartzite layers.

UPPER SANDY MEMBER OF THE GATESBURG FORMATION

	Feet
Covered interval with typical weathered sandstone blocks. Measurements begin at fartherest east block in place	456
65. Buff silty dolomite, massive to platy-bedded	5
64. Buff quartzite	1
63. Massive dark dolomite	1
62. Covered interval	4.5
61. Platy buff silty dolomite	4
60. Massive dark dolomite	8
59. Platy buff silty dolomite	2
58. Thin-bedded blue-weathering limestone	4
57. Dark-buff silty dolomite	2
56. Shaly, platy, thin-bedded dolomite	2.5
55. Hard massive dark-gray dolomite	3
54. Platy bluish limestone	2
53. Massive gray dolomite	6
52. Quartzite with buff silty dolomite at top	5
51. Massive dolomite, laminated and light-colored at top	17
50. Massive dark limestone	6
49. Dark thin-bedded limestone with platy dolomite seams	2
48. Dark aphanitic cherty limestone with cryptozoon and pebble-conglomerate	5
47. Massive sandy buff dolomite, cherty at top	2
46. Dolomitic quartzite	2.5
45. Light-colored buff silty dolomite and beds of quartzite	3.5
44. Buff silty platy, thin-bedded, laminated dolomite whose top is capped with thin quartzite stringers	12
43. Black aphanitic limestone, black nodular chert, cryptozoon	5
42. Dolomite weathering to irregular masses, sandy with cherty oölite	3
41. Quartzite grading up into crystalline dolomite	1
40. Black massive dolomite weathering to ribby surfaces	1.5
39. Shaly dolomite	1
38. Black finely crystalline buff-weathering dolomite	1
37. Platy-weathering limestone with yellow dolomite mottling in layers	4.2
36. Massive hard black aphanitic limestone, weathering bluish	5.5
35. White massive crystalline dolomite, cryptozoon throughout capped by black crystalline dolomite	3.5
34. Black aphanitic massive hard pure limestone	4.5
33. Massive finely crystalline buff silty dolomite with half a foot of quartzite	6
32. Covered interval	8
31. Aphantic laminated dolomite with shaly bed at top	2.5
30. Sandy dolomite and some quartzite grading into pure dolomite	3.5
29. Dark finely oölitic crystalline dolomite, finely cross-bedded and sandy at top	7

28. Pinkish massive, finely crystalline dolomite and coarse quartzite	2.5
27. Mostly covered, but platy, silty-sandy beds	4
26. Buff silty massive dolomite	4.5
25. Silty massive irregularly laminated mottled limestone, dolomitic at base	8
24. Dark-gray limestone capped by finely developed cryptozoon 47-11w.3	3
23. Dark finely crystalline dolomite, weathering buff and bearing cryptozoon, capped by pink dolomite conglomerate with thin platy type pebbles 47-11w.3	2.7
22. Dark crystalline dolomite	3.5
21. Dark but light-weathering massive dolomite with horny chert and cryptozoon	4
20. Thin-bedded quartzite	2.5
19. Black massive finely crystalline dolomite	11.5
18. Black crystalline limestone	.5
17. Pink dolomite conglomerate with flat, thin pebbles, possibly *Hyolithes*	3.5
16. Laminated pinkish-blue bed of silty dolomite	.5
15. Massive dolomite, soft weathering into ribby masses	7
14. Pink dolomitic limestone, possibly with *Hyolithes*	2
Total well-exposed upper sandy member of the Gatesburg	217
Total upper sandy member of Gatesburg	673

THE ORE HILL MEMBER OF THE GATESBURG

(The position of its top is arbitrarily drawn)

13. Black limestone weathering mottled with yellow spots	4
12. Black finely crystalline massive dolomitic limestone	2
11. Black finely crystalline to aphanitic limestone weathering platy and yellow mottled. Middle of this interval is 47-11w.4, *Taenicephalus*, primitive orthoid brachiopod	11.5
10. Black pebble conglomerate with crystalline matrix grading up into limestone	1
9. Black finely crystalline to aphanitic limestone weathering steel blue and with yellow mottling. 47-11w.2, float at approximately this position, primitive orthoid, *Taenicephalus*	2
8. Dark-gray very finely crystalline dolomite grading into white very pure aphanitic to finely crystalline massive limestone for top 1½ feet	9.5
7. Massive crystalline dolomite, buff white, some thin-bedded and laminated dolomite	6
6. Mostly covered, top foot is thin-bedded silty dolomite	6
5. Massive gray crystalline dolomite	2.5
4a. Poorly exposed, thin-bedded limestone float with a pink pebble conglomerate, flaggy limestone and cryptozoon	10
4b. Covered interval	68
3. Crystalline dolomite	2
2. Covered interval	21
1. Dark aphanitic limestone weathering ribby and irregular	5
Total exposed Ore Hill member	150.5

LOWER SANDY MEMBER OF THE GATESBURG FORMATION

Covered interval	307
Quartzite	1
Covered interval	6
Thin-bedded sandy dolomite and quartzite	2
Conglomeratic quartzite	1.5
Soft-weathering quartzite	3.5
Dark massive crystalline dolomite	3
Coarsely sandy light-colored dolomite	.5
Chippy-weathering light-colored dolomite	1
Coarse gray quartzite	2
Cross-bedded arenaceous dolomite	1
Buff silty dolomite	.5
Massive dark crystalline dolomite, oölitic at top	2
Total well-exposed lower Gatesburg	24

FOLLOWING BEDS WRINKLED INTO SMALL FOLDS

Massive dolomite at base; top covered	29
Covered interval	28
Massive dolomite	4
Quartzite	5
Massive dolomite	6
At base is cherty, chippy-weathering dolomite much like certain Stacy outcrops; some weathered quartzite present above this	75
Buff silty dolomite	1.5
Covered interval	18.5
Massive dolomite	4
Quartzite	1
Crystalline dark dolomite with some buff silty dolomite beds up to 2 feet thick	79
Covered interval	7
Nodular lumpy-weathering dolomite and some covered beds; at top is buff silty-sandy platy dolomite with quartzite layers	16.5
Arenaceous dolomite and quartzite, thin-bedded dark dolomite	4
Hard quartzite	2.5
Soft platy quartzite and dolomite, mostly covered	20
Total lower sandy member of the Gatesburg	632

STACY DOLOMITE MEMBER OF THE GATESBURG

Massive hard dolomite	17
Soft-weathering thin-bedded dolomite	13
Covered interval	26
Massive dolomite	3
Covered interval	24
Total Stacy member of the Gatesburg	83
Total Gatesburg formation exposed	1538.5

WARRIOR FORMATION

Shaly limestone	7
Covered interval	77
Aphanitic limestone or dolomite which weathers white	2.5
Covered interval	3.5
Massive cryptozoon limestone	19
47-11w.1, *Lonchocephalus swartzi, Kingstonia apion, Meteoraspis, Maryvilla* sp. undet.	
Covered interval, limestone	20
Thin-bedded platy limestone	11
Additional section in the Warrior measured in drainage ditch on north side of Turnpike west of low ridge cut through by the road and exposing part of Warrior given above. Dip 15° S. 75° E.	
Covered interval	7.5
Mostly thin-bedded limestone, crystalline and fossiliferous	18
Covered interval	11
Platy limestone	3
Covered interval	14
More massive limestone with cryptozoon at intervals	19
Platy dolomite	2
Cryptozoon limestone	4
Dolomite	9
Massive dolomite with cryptozoon near base	11
Covered, probably sandy siltstone	10
Massive dolomite	10
Total exposed Warrior formation	258.5

Detailed Sections of the Ore Hill Member

LOCALITY 47-3w, Drab: Fossils were collected here by Butts and Ulrich on wooded slopes north of and in a low road cut at the south side of an asphalt township road a quarter of a mile west of Drab-Beavertown in the northern part of Huntingdon quadrangle, Pennsylvania. Dips are steep enough (30°–40°) so that the beds strike steeply up the slope and may be traced through the bushes with some difficulty. A complete section exposing both contacts was obtained by continuing across the road. The highest beds are in a small quarry just east of the road cut on the south side. The beds in the whole section are very fossiliferous and represent the *Pseudosaratogia magna* fauna and the typical *Elvinia* zone. The *Conaspis* fauna was not found.

TOPMOST BEDS OF THE ORE HILL MEMBER OF THE GATESBURG: (SOUTH OF ROAD)

	Feet
47-3w.36, hard gray aphanitic to finely crystalline, laminated, massive limestone	top
47-3w.35, steel-gray, fine- to medium-crystalline massive dolomitic limestone	6.5
47-3w.34, dark aphanitic laminated limestone	7.0
Covered interval	4.0
47-3w.33, dark platy, finely crystalline to aphanitic hard limestone	4.0
47-3w.32, dark crystalline massive limestone	3.0
47-3w.31, dark crystalline massive limestone grading into finely crystalline lighter-colored laminated dolomitic limestone	4.0
47-3w.30, dark crystalline dolomitic limestone with undulating lamination	2.0
Covered interval	6.5
47-3w.29, dark hard finely crystalline limestone, *Pinctus latus, Bultsia drabensis, Pseudosaratogia* sp. (top beds exposed in road cut. Probably an environmentally controlled biofacies different from typical Elvinia fauna)	
Black aphanitic to finely crystalline limestone weathering to whitish thin beds or massive beds all covered with hackly markings producing almost a honeycomb effect	14.5
47-3w.28, coarsely crystalline limestone flat pebble conglomerate, matrix gray, pebbles pinkish and aphanitic, fossiliferous, *Bultsia drabensis, Burnetia urania, Camaraspis convexa, Cliffia lataegenae, Drabia acroccipita, Deckera completa, Dellea suada, Deadwoodia duris, Irvingella major, Pinctus latus, Pseudosaratogia bulbosa*, oboloid brachiopod	.5
(Section now continues down on north slopes of hill across road)	
47-3w.27a, hard aphanitic to finely crystalline limestone	5.0
Covered interval, probably like unit above	7.5
47-3w.27, black aphanitic limestone, weathering white, nodular or platy	4.0
47-3w.26, black aphanitic limestone as above, *Bultsia drabensis, Bynumina terrenda, Burnetia urania, Cliffia lataegenae, Dellea suada, D. bullerensis, Deckera completa, Deadwoodia duris, Drabia acroccipita, Elvinia roemeri, Housia vacuna, Irvingella major, Kyphocephalus ponderosus, Kindbladia wichitaensis, Platyaspella anatina, Pinctus latus, Pseudosaratogia bulbosa, Xenocheilos spineum*, pyg. B, oboloid brachiopods	3.0
Covered interval	4.0
47-3w.25, dark aphanitic to crystalline limestone, trilobite fragments	4.5
47-3w.24, limestone pebble conglomerate, coarsely crystalline matrix, pebbles are pink and aphanitic	.5
47-3w.23, black aphanitic limestone weathering white and platy, covered with hackly markings	5.5

Elvinia faunizone

Elvinia faunizone

47-3w.22, coarsely crystalline trilobite coquina, *Burnetia urania, Bynumina terrenda, Cliffia lataegenae, Deadwoodia duris, Drabia acroccipita, Dellea butlerensis, Dellea saratogoensis, Deckera completa, Elvinia roemeri, Housia vacuna, Irvingella major, Kyphocephalus ponderosus ?, Platyspella anatina,* oboloid brachiopod 5

47-3w.21, dark finely crystalline aphanitic limestone, weathers with hackly markings 8.0

47-3w.20, gray crystalline limestone with some flat pink pebble conglomerate, *Buttsia drabensis, Burnetia urania, Bynumina terrenda, Cheilocephalus buttsi, Cliffia lataegenae, Deadwoodia duris, Deckera completa, Dellea suada, Elvinia roemeri, Housia vacuna, Irvingella major, Kindbladia wichitaensis, Platyspella anatina ?, Pseudosaratogia lata, Xenocheilos spineum,* oboloid brachiopods 5

Covered interval 5

47-3w.19, coarsely crystalline gray trilobite coquina zone only an inch thick, *Buttsia drabensis, Burnetia urania, Bynumina terrenda, Cheilocephalus buttsi, Cliffia lataegenae, Deadwoodia duris, Dellea butlerensis, Deckera completa, Elvinia roemeri, Housia vacuna, Irvingella major, Kyphocephalus ponderosus, Kindbladia wichitaensis, Platyspella anatina?, Pseudosaratogia lata, Xenocheilos spineum,* oboloid brachiopods.

47-3w.18, pink pebble conglomerate, gray matrix, coarsely crystalline, *Bynumina terrenda, Dellea* sp., *Deadwoodia duris, Elvinia roemeri, Irvingella major, Kindbladia wichitaensis* 5

Covered interval 4.0

47-3w.17, coarsely crystalline gray trilobite coquina, *Bynumina terrenda, Cliffia lataegenae, Deadwoodia duris, Deckera completa, Dellea suada, Irvingella major, Kindbladia wichitaensis* .5

47-3w.17a, dark aphanitic limestone, nodular weathering 2.5

47-3w.16, pink pebble conglomerate, gray crystalline limestone matrix, *Cliffia lataegenae, Deadwoodia duris, Dellea* sp., *Irvingella major, Housia vacuna* 5

Covered interval 1.0

47-3w.15, gray coarsely crystalline limestone, *Burnetia urania, Bynumina terrenda, Dellea suada, Dellea butlerensis, Cliffia lataegenae, Elvinia roemeri, Housia vacuna,* pyg. B. *Hyolithes* 5

47-3w.14a, crystalline limestone with great abundance of reddish brown *Hyolithes, Drabia acroccipita, Housia vacuna* 3.0

47-3w.14, dark crystalline limestone with abundance of *Hyolithes;* lowest occurrence of typical *Elvinia* fauna, *Cliffia lataegenae, Drabia acroccipita, Dellea saratogoensis, Elvinia roemeri, Housia vacuna,* pyg. B 3.0

Elvinia faunizone

Covered interval 6.5

47-3w.13, dark crystalline limestone 5

47-3w.12, dark, finely crystalline limestone 3.5

47-3w.11, light-gray dolomitic limestone 3.5

Covered interval 9.0

47-3w.10, dark limestone, dolomitic, weathers light 5

47-3w.6, float, from above between 47-3w.14 and 3w.10; typical *Elvinia* fauna with probable highest occurrences of the *Pseudosaratogia magna;* also *Buttsia drabensis, Berkeia granulosa, Cliffia lataegenae, Bynumina terrenda, Drabia acroccipita, Dellea butlerensis, Deckera completa, Irvingella major, Kyphocephalus ponderosus, Pseudosaratogia magna, Hyolithes*

Elvinia faunizone

Covered interval 7.5

47-3w.5, gray crystalline dolomite 1.0

47-3w.4, gray crystalline oölitic limestone or dolomite 1.0

Total Ore Hill member of the Gatesburg 144.0

Lower sandy member of the Gatesburg: pure quartz sandstone and medium crystalline dark oölitic limestone 2.5

LOCALITY 47-5w, Potter Creek-Scott Fisher Farm: Along the outcrop exposed on the north side of Potter Creek and behind (north of) the farm house belonging to Scott Fisher. The beds are traceable from their lowest exposures northwest of the farm house along the ridge for about a quarter of a mile east until the upper part of the Ore Hill is cut by the road following the creek from Maria (Pennsylvania State Highway 868), northwest Everett quadrangle, Pennsylvania.

ORE HILL MEMBER OF THE GATESBURG FORMATION

	Feet
47-5w.44, gray platy-weathering dolomite with undulating laminae	
Covered interval, probably as above ...	3.0
47-5w.43, massive steel-gray finely crystalline dolomite	6.0
47-5w.42, dark-gray, finely crystalline platy limestone with trilobite fragments	3.0
47-5w.41, finely crystalline massive dolomitic limestone	4.0
47-5w.40, dark-gray finely crystalline platy limestone	2.0
47-5w.39, aphanitic to finely crystalline dark nodular-weathering limestone ...	2.0
47-5w.38, very finely crystalline laminated light-weathering platy limestone	1.5
47-5w.37, dark aphanitic pebble conglomerate, finely crystalline matrix ..	2.0

←*Conaspis* faunizone?

This section is almost complete and is the first well-exposed and fossiliferous section south of Drab, 47-3w. It is complemented by section 47-6w south of the creek. It furnished a large part of the excellent trilobite material from the *Elvinia* and *Conaspis* zones.

The above beds were measured in and just above the road cut exposing the upper Ore Hill member at the east end of its outcrop, north of Potter Creek. The following collection (47-5w.36) is located high on a bluff behind an old house south of the creek opposite the road cut. Its position in the section was obtained by measuring down from the base of the dolomite beds at the top of the member. It is a thin bed and is fitted into the section at its approximate position. It is not a thickness unit.

47-5w.36, dark aphanitic pebble conglomerate with crystalline matrix, *Conaspis perseus?*, *Wilbernia* sp., *Taenicephalus shumardi*.

(Section returns to north of creek)

Conaspis faunizone

47-5w.35, medium-crystalline nodular dark limestone	1.5
47-5w.34, light-gray coarsely crystalline limestone pebble conglomerate with trilobite fragments	4.0
Base of this bed is a trilobite coquina of finely crushed free cheeks with a few identifiable cranidia; the matrix is coarsely crystalline, and limestone pebbles are also present, *Conaspis perseus*, *Conaspis(?) retractalabra Taenicephalus shumardi*, *Pelagiella* sp.	.5
47-5w.33, dark aphanitic limestone, platy and light-weathering, sample toward top	6.5
47-5w.32, coarsely crystalline limestone with trilobites and a pebble conglomerate with tiny black grains, massive weathering	5.5
47-5w.31, massive finely crystalline limestone (forms ledge)	1.0
Covered interval	2.0
47-5w.30, dark finely crystalline to aphanitic platy limestone	3.0
47-5w.29, coarsely crystalline limestone with trilobite fragments	3.0
47-5w.28, coarse pink and dark aphanitic pebble conglomerate, crystalline matrix, possibly inaccurately placed in section	.5
47-5w.27, finely crystalline massive gray limestone	2.5
47-5w.26, dark finely crystalline massive to platy limestone	1.5
47-5w.25, dark aphanitic to finely crystalline laminated limestone	1.5
47-5w.24, dark aphanitic to finely crystalline laminated limestone	1.5
47-5w.23, dark finely crystalline limestone with trilobite fragments	1.0
47-5w.22, dark finely crystalline light-mottled weathering limestone	2.5
47-5w.21, dark crystalline platy limestone	1.0
47-5w.20, dark crystalline platy limestone	1.5
47-5w.19, dark platy crystalline limestone with black nodules and trilobite fragments	1.5
47-5w.18, dark well-laminated crystalline limestone	2.0
Covered interval	1.0
47-5w.18a, coarsely crystalline limestone with primitive orthoid brachiopod	1.0
47-5w.17, finely crystalline dark limestone, primitive orthoid, *Parabolinella occidentalis*, *Wilbernia* sp.	5.5
47-5w.16, dark crystalline nodular-weathering limestone	1.5
47-5w.15, No sample	
47-5w.14, dark crystalline limestone, finely and strikingly laminated, weathers sharply platy	4.5
47-5w.13, dark finely crystalline limestone	4.0
47-5w.12, dark finely crystalline limestone with conaspid trilobite	
Dark finely crystalline aphanitic limestone, 47-5w.11 at base	5.0
Covered interval	16.0
47-5w.10, dark aphanitic platy limestone	? .5
Covered interval	10.0
47-5w.9, massive to platy aphanitic limestone	3.5

Elvinia faunizone

47-5w.8a, crystalline limestone, pebble conglomerate, highest occurrence, *Berkeia granulosa*, *Buttsia drabensis*, *Burnetia urania*, *Cliffia lataegenae*, *Camaraspis convexa*, *Camaraspoides berkeyi*, *Deckera completa*, *Dokimocephalus intermedius*, *Dellea suada*, *Dellea saratogoensis*, *Drabia acroccipita*, *Drabia minuta*, *Elvinia roemeri*, *Pterocephalia ' sanctisabae*, pyg. B, *Cheilocephalus*, pyg. L, oboloid brachiopods, *Hyolithes*.	
47-5w.8, dark aphanitic nodular and platy-weathering limestone	1.0
47-5w.7, dark aphanitic platy limestone	2.5
Covered interval	9.0
47-5w.6, dark aphanitic platy limestone	3.0
47-5w.5a, dark aphanitic platy limestone	3.5
47-5w.5, trilobite coquina in pink pebble conglomerate, *Buttsia drabensis*, *Berkeia granulosa*, *Burnetia urania*, *Bynumina terrenda*, *Camaraspis convexa*, *Cliffia lataegenae*, *Deckera completa*, *Deadwoodia duris*, *Drabia acroccipita*, *Drabia minuta*, *Dellea suada*, *Dellea saratogoensis*, *Elvinia roemeri*, *Irvingella major*, *Housia vacuna*, *Pinctus latus*, oboloid brachiopods, *Hyolithes*.	.5
Black aphanitic limestone	3.0
47-5w.3, gray crystalline trilobite co-	

Elvinia fauna zone	quina with pink pebble conglomerate; *Buttsia drabensis, Burnetia urania, Camaraspis convexa, Cliffia lataegenae, Dokimocephalus intermedius, Dellea suada, Dellea saratogoensis, Deckera completa, Drabia acroccipita, Elvinia roemeri*, oboloid brachiopods	.5
	Black aphanitic limestone	5.5
	47-5w.3a, crystalline limestone, dark with black aphanitic nodules, platy	4.0
	47-5w.2, black aphanitic limestone	2.5
	47-5w.1, black aphanitic limestone	
	Base cut out by fault	
	Total Ore Hill member exposed	155.0

LOCALITY 47-20w, Imler Quarry: Just off and east of township road east of Evitt's Mountain, northwest corner of Clearville quadrangle (Colerain township) at a point 2.5 miles south of where the road leaves the Bedford-Rainsburg road at Egolf Park. Section lies in the west face of a wooded ridge 200 yards east of road. Dip about 9° from S. 10° W. to S. 30° W. The lower part of the section lies in a small abandoned quarry near the foot of the ridge. The upper portion lies in low rock ledges at the crest of the wooded ridge.

ORE HILL MEMBER OF THE GATESBURG FORMATION

		Feet
Conaspis zone	Dark platy limestone forming a small cliff	5.0
	47-20w.2, *Conaspis perseus*, oboloid brachiopod	
	Dark platy limestone forming a low cliff	15.0
	Covered interval	1.0
	47-20w.6, *Taenicephalus shumardi, Pelagiella* sp.	
	Platy dark aphanitic limestone	2.0
	47-20w.5, *Taenicephalus shumardi, Pelagiella* sp.	
	Platy dark aphanitic limestone	1.0
	Massive white limestone, some dolomite grading up into dark limestone as above	27.0
	Silty dark dolomite	7.0
	Massive white limestone	6.0
	Massive white limestone and dolomite	22.5
	Covered with float, probably massive nonresistant dolomite	15.5
	Massive and dark crystalline dolomite	6.0
	Massive dolomitic limestone (top of quarry)	5.0
Elvinia zone	Thin-bedded platy limestone	1.5
	Massive dark crystalline limestone, worm-eaten appearance	1.5
	Massive dark crystalline limestone, 47-20w.3, *Pseudosaratogia bulbosa*.	
	Dark thin-bedded limestone	2.0
	47-20w.1a, pink limestone pebble conglomerate, *Burnetia urania*	
	Dark thin-bedded limestone	2.0
	47-20w.4, *Hyolithes*? bed in woods north of quarry	
	Dark thin-bedded limestone	8.5
	47-20w.1, crystalline limestone exposed in woods just north of quarry and may be traced into floor of quarry, *Buttsia drabensis, Cliffia lataegenae, Berkeia granulosa, Dellea suada, D. saratogoensis, Deadwoodia duris, Deckera completa, Irvingella major, Pinctus latus, Pseudosaratogia bulbosa.*	
	Thin-bedded platy dark aphanitic limestone	12.0
	Total exposed Ore Hill member of the Gatesburg	141.0

(A comparison of this section with that from Fickus quarry—*see* Wilson, 1951, p. 621—indicates that at least 25 feet of beds is unexposed below the lowest beds measured at Imler quarry; the *Pseudosaratogia magna* fauna is not reached at the base of this section.)

FORMATIONS IN THE CUMBERLAND-SHENANDOAH VALLEYS

General Discussion

No Cambrian formations are exposed across the southern Pennsylvania mountains immediately east of Nittany Arch except in one anticline in the western Mercersburg quadrangle (Stose, 1909). This area was mapped by Stose as Beekmantown, but considerable late Upper Cambrian limestone is exposed along the axis of the structure forming McConnellsburg Cove. The main exposures of Upper Cambrian formations in this area lie in the Cumberland Valley and its southern continuation, the Shenandoah. Relatively nonresistant limestone of this age on the west flank of South Mountain anticlinorium forms the eastern side of the valley. The Massanuten syncline trends north-south about the middle of Cumberland Valley preserving the Martinsburg shale and underlying Ordovician limestone. A western anticline in the valley again brings Cambrian beds to the surface in Maryland and West Virginia, but this plunges northward, and beds this old are not known in southern Pennsylvania north of the west branch of Conococheague Creek (Fig. 1). A reverse fault, probably an extension of (Little) North Mountain fault Butts (1940, p. 452), cuts off the west flank of this fold and forms the western boundary of the valley from Frederick County, Virginia, north into southern Pennsylvania.

This area is covered on maps by Stose (Cham-

bersburg-Mercersburg quadrangles, Pennsylvania, 1909), Cloos (Washington County, Maryland, 1941), Grimsley (Jefferson, Berkeley, and Morgan Counties, West Virginia, 1916), and Butts and Edmundson (manuscript, Frederick County, Virginia).

Cambrian formations examined are those above the Lower Cambrian Waynesboro and include the Middle-Upper Cambrian Elbrook shale and dolomite and the Upper Cambrian Conococheague limestone. These two formations aggregate about 5000 feet thick.

Elbrook Formation

General statement.—The Elbrook formation was named by Stose (1906, p. 209) for a village north of Waynesboro, Pennsylvania, prior to publication of the Mercersburg-Chambersburg folio (Stose, 1909). The writer has examined the upper part of the formation 2 miles south and west of Waynesboro, Pennsylvania (47-25w), in Antietam Battlefield (47-27w), and at Big Spring Station (47-29w), Maryland, and collected fossils only from the upper 700 feet.

Lithologic details.—Stose (1909, p. 5) says that the upper half of the formation is "composed largely of light-colored calcareous shale and laminated impure limestone which weathers shaly." This rock may be better described as yellow-weathering dolomite and calcareous dolomitic shale darker when fresh but weathering light. These beds are finely laminated or thin-bedded and commonly alternate with limestone. One distinctive type of Elbrook limestone is light, pure, and finely crystalline and weathers a pastel pink or blue. The only insoluble residue obtained from this was a minute quantity of fine white sand. Gray mottled limestone of this general type may be interbedded with the Elbrook shale. The formation also contains dark, finely crystalline, slate-blue weathering silty laminated and generally fossiliferous limestone exactly like that of the Conococheague formation. These lithic units seem to have no regular succession in the area studied. The Conococheague-type limestone has been noted from the very top of the Elbrook to as low as 800 feet below the base of the Conococheague.

Geographic expression and variation along strike.—The Elbrook is much less resistant than the overlying Conococheague formation and in the Chambersburg quadrangle forms lower, flatter grassy land with almost no outcrops. However, in Maryland southward from the type locality the upper Elbrook becomes less argillaceous, more calcareous and dolomitic, and its light limestones form ridges across the fields just as do those of the Conococheague formation. Some hundreds of feet of steeply dipping calcareous beds are well exposed in Antietam Battlefield west of West Woods. Grimsley's measured section (1916, p. 292–301) of the upper Elbrook exposed along the Potomac east of Shepherdstown, West Virginia, shows about 1000 feet of beds none of which is described as shale. In part this southward change in lithofacies may have caused Cloos to pick a different position for the Conococheague-Elbrook contact in Washington County, Maryland, from that designated by Stose at the type locality. Cloos (1941, map) includes all the upper Elbrook in the Conococheague. (*See* discussion of Conococheague sandstone.)

Butts (1940, p. 74–79) describes the Elbrook dolomite of southern Virginia in some detail but little is known of the formation in northern Virginia. Stose (1909, p. 6) states that the Elbrook is 3000 feet thick but it thins southward to about 1500 feet (Butts, 1940).

Age of the Elbrook formation.—Resser (1938, p. 22, 79) notes that the basal beds of the formation near Waynesboro have yielded a species of *Glossopleura* also found in the Rutledge of Tennessee. This places the base of the formation as rather low in the Middle Cambrian sequence. However, much of the upper 700 feet of the Elbrook may be of Dresbachian age.

The faunas of the Warrior formation are very similar to those of the Big Spring member of the Conococheague and the uppermost Elbrook formation. Beds 75 to 200 feet below the top of the Big Spring Station member contain an undoubted *Crepicephalus* zone assemblage, although the name genus of the faunizone is lacking. Pygidia of *Pemphigaspis* and *Llanoaspis* have been found, however, and, as in the Warrior formation, *Tricrepicephalus* seems confined to this stratigraphic level. Several of the common Dresbachian genera range down into the Elbrook below the base of the Conoco-

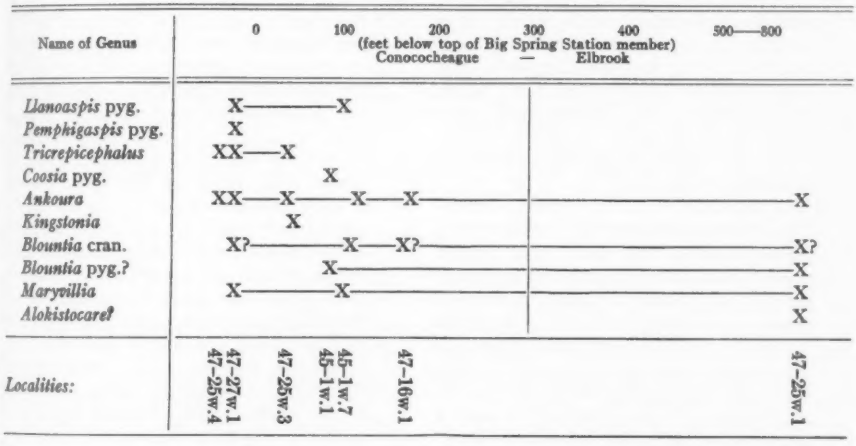
cheague and are found about 800 feet below the top of the Elbrook at one locality. Here they are associated with a new and small species of the Middle Cambrian trilobite *Alokistocare*. This fauna (47-25w.1) represents either the *Cedaria* faunizone of the standard Dresbachian or high Middle Cambrian.

Butts's sections (1940, p. 74–76) of the upper Elbrook dolomite of southern Virginia list several well-known Dresbachian trilobites, and the writer considers that the Middle-Upper Cambrian boundary falls toward the top of this formation.

ing, and edgewise conglomerate are of value. Cloos (1941, map) shows considerable structural detail in these beds in the Cumberland Valley, Washington County, Maryland, but assumes different boundaries for the Conococheague from those set by Stose's original work. (*See also* Cloos, 1951, p. 45–57.) In West Virginia, these beds have not been studied since Grimsley's report in 1916. Butts's work in Virginia is important (1940, p. 86–90).

The writer's study of 12 measured sections of the Conococheague reveals that the formation consists of about 2000 feet of dark silty

TABLE 4.—TRILOBITE OCCURRENCES IN THE ELBROOK FORMATION AND BIG SPRING STATION MEMBER OF THE CONOCOCHEAGUE



Conococheague Limestone

Description.—Beds now known as the Conococheague limestone were originally included in the Knox formation (Stose, 1906, p. 209, 210). At that time the distinctive basal sandy beds of the present Conococheague were first described but not named. Later (1908, p. 701–703; 1909, p. 6) Stose described the lower part of the Knox in more detail, naming it the Conococheague limestone from exposures along the creek of that name north of Chambersburg, Pennsylvania.

Bassler (1919, p. 74–88) describes the Maryland Conococheague exposures south of the Chambersburg quadrangle. His photographs and descriptions of cryptozoon, siliceous band-laminated slate-blue limestone which is lighter toward the top and possesses a basal sandy member here designated the Big Spring Station member.

Sections studied are as follows:

Southern Pennsylvania:

Scotland section. (Stose, 1909, p. 6) West side of Scotland, Chambersburg quadrangle. About upper two-thirds of formation.

47-23w: Musser Farm. East side of Scotland, Chambersburg quandrangle. Lower 500 feet of formation.

47-9w: North of Stonehenge. Inlier on Stose's map of Chambersburg quadrangle, 1½ miles north of Stonehenge village. Upper Conococheague only.

47-12w: Grindstone Hill. Big Spring Station member forms hill.

47-26: Waynecastle Dairy. 3½ miles east of Greencastle at Western Maryland Railroad crossing of Waynesboro road. Almost a complete section of formation exposed in road cuts.

47-25w: South of Waynesboro. 2 miles south of Waynesboro where Hagerstown road crosses junction of branches of Antietam Creek.

Maryland, Washington County:

Eastern belt:

47-28w: Roxbury Mills. West from Antietam Creek crossing at old abandoned mills and across hills to penal farm of same name. (*See* Cloos, 1951, p. 46-52, for a detailed measured section here). Both the writer's and Cloos's sections begin at a fence south of road across from an old house and proceed in the same direction. Beds of the Big Spring Station member are about 300 feet thick and are only slightly sandy but have considerable dolomite. About 600 feet above the base of the section occur trilobites of the *Pseudosaratogia magna* fauna, the only fauna encountered in the section. *P. magna*, *Buttsia drabensis*, and *Drabia curtoccipita* (47-28w.2).

47-27w: Antietam Battlefield. North of Sharpsburg, Maryland. About 1000 feet of Elbrook exposed west of West Woods; silty resistant dolomite beds of the Big Spring Station member support the ridge east of Hagerstown Pike and cross same at Miller's Cornfield. About 500 feet of the lower Conococheague limestone is exposed west of the axis of a syncline which crosses Mansfield Avenue, 200 yards east of the pike. Fossils at the top of this interval north of Mansfield Avenue include the highest *Pseudosaratogia magna* collected in the Conococheague; also *Stigmacephalus ? distorta* (47-27w.2, Wilson, 1951).

Maryland, Washington County:

Western belt:

47-16w, Cushwa Mill: ½ mile southwest of Cushwa Mill, just south of a branch of Rockdale Creek and south of a series of farm houses north of the creek branch, north of Clear Spring, central Williamsport quadrangle. (*See also* Wilson, 1951.) Interpretation of geology in this area differs considerably from that of Cloos (1941) and Stose (manuscript map). Structure is obscure, but discovery of a fine Franconian trilobite fauna at the expected interval above the Big Spring Station member clarifies the stratigraphy. (About 240 feet of the Big Spring Station member and almost 200 feet of middle Conococheague limestone above it are exposed.)

47-29w, Big Spring Station. Railroad cuts north of and in fields west of Big Spring Station on the Western Maryland Railroad about 3 miles south of Clear Spring. The section is almost complete. The railroad cuts expose about 50 feet of Elbrook and 312 feet of very arenaceous Big Spring Station member overlain by about 180 feet of dolomitic shales and limestones and followed above by 630 feet of typically laminated Conococheague limestone. The hill between the automobile road and the railroad and the area west of the road behind Mr. Newkirk's house and barn continue the higher parts of the section adding about 800 feet of limestone containing Trempealeauian fossils in its middle portion. This section was published for the writer by E. Cloos (1951, p. 53–57).

Virginia, Frederick County:

Western belt:

45-1w, Winchester. Part I, in a meadow along a railroad spur 2 miles south of the heart of the city. Upper portion exposed in an apple orchard west of sheet-metal apple warehouse near a large red brick processing plant. Section directed due west for 1800 feet to crest of wooded ridge formed on Big Spring Station member. A complete but poorly exposed section showing both lower and upper Conococheague sandstones and fossils of Trempealeauian and Franconian age. About 1800 feet of Conococheague limestone above the lower member. Part II (*See also* Wilson, 1951), on a low hill one quarter of a mile west of city center just south of city limit sign on U. S. Route 50. Hill part of same ridge in Part I supported by 370 feet of Big Spring Station member. These beds contain Dresbachian trilobites. Top of this interval coincides with the uppermost sandstone on east flank of hill. This is overlain by 275 feet of well-exposed typical Conococheague limestone, the upper 130 feet bearing a Franconian trilobite fauna. Upper part of the section lies near a rock wall west of Glen Burnie estate.

Type section and thickness of the Conococheague formation.—A section of the formation 1650 feet thick measured by Stose just west of the village of Scotland on Conococheague Creek, 3 miles northeast of Chambersburg, Pennsylvania, became unofficially the type section, although the section is incomplete, and its lower portion is structurally distorted.

A minor anticlinal axis passes through the village exposing the lower part of the Conoco-

cheague on both its flanks. The beds across the axis are covered, and the section cannot be traced across it at Scotland. Stose overestimated the height of the fold, believing that the underlying Elbrook formation was exposed along it south of Scotland (1909, map). Fossils of Trempealeauian age (high in the Conococheague formation) discovered within this supposed Elbrook inlier 1½ miles north of Stonehenge village indicate that the structure is relatively small. Therefore, on the west limb of this minor fold at Scotland, less of the lower Conococheague is repeated than Stose believed, and his estimated total thickness is too small To offset this, Closs (1951, p. 46) points out that crenulated laminations and deformed oölites measured in this section show that the beds measured here are overthickened structurally.

About 500 feet of the lower Conococheague (a part of Big Spring Station member at the base) is present on the east flank of the fold at Musser Farm (47-23w), on the east edge of Scotland village. These beds are not present in Stose's section.

The Waynescastle Dairy section (47-26w), the best exposed section of the Conococheague, is 1890 feet thick using the boundaries drawn by Stose on the map of Chambersburg quadrangle. The Big Spring Station section (47-29w) is almost complete and about 1900 feet thick (Cloos, 1951, p. 53–57). At Winchester, Virginia (45-1w) the section according to Butts's mapping includes 2170 feet. Butts (1940, p. 89) gives about 2000 feet as the thickness of the Virginia Conococheague.

Big Spring Station member of the Conococheague.—Bassler (1919, p. 78) divides the Conococheague of Maryland into three main units: (1) the lower 250 feet of "oölite, edgewise conglomerate and Cryptozoon reefs"; (2) the thick middle portion of typically laminated limestone (about 1000 feet); and (3) an upper 400 feet which includes "pink marbles" in addition to the banded limestone. The lowest unit is persistent but not necessarily characterized by the rock types listed by Bassler. It is actually a dolomite-orthoquarzite unit and is here named the Big Spring Station member. The other two units of Bassler are not distinct enough to warrant separation and naming.

The writer proposes a formal name for the lower unit of the Conococheague to emphasize its importance both in areal mapping and in stratigraphic correlation. The name is taken from the excellent and complete exposures of the member at its thickest in the railroad cuts at Big Spring Station. (*See* section in Cloos, 1951, p. 53–57, published for writer but under incomplete name Big Spring member.) These beds were first recognized by Stose who noted their more clastic content and suggested erroneously that they represented the base of the Upper Cambrian.

Six criteria considered by Stose (1908, p. 701–703; 1909, p. 6) as distinctive of these lower beds and emphasized by Bassler (1919, p. 78, 80) as indicative of shallow-water deposition are: (1) more siliceous; (2) rounded limestone pebble conglomerates with a matrix of coarse sand grains; (3) edgewise conglomerates; (4) predominance of oölites; (5) predominance of cryptozoon; and (6) dark shaly limestone with red clay inclusions. The writer believes these criteria need modifications. Stose's phrase "more siliceous" is considered to mean that the beds are more arenaceous. The edgewise conglomerate, cryptozoon, and oölitic limestones mentioned as characteristic of the member are not confined to it although present in many of its sections. This is particularly true of the cryptozoon stromatolite. The writer found no trace of the red clay inclusions mentioned by Stose and Bassler and considers them a weathering phenomenon. The Big Spring Station member is best defined by abundant beds of shaly or platy yellow-weathering dolomitic siltstone or silty dolomite and coarsely arenaceous dolomite or siliceous orthoquartzite. Some of the latter beds are from 4 to 5 feet thick, although generally the rock type occurs in thinner layers interbedded with the typical laminated limestone of the higher Conococheague. Much of the limestone is oölitic or composed of edgewise intraformational conglomerate as noted above. These sandy beds of the lower Conococheague weather into porous yellow blocks which litter the surface (Stose, 1909, p. 6) and support a wooded ridge wherever they crop out. The member is thus important as a key bed for mapping the nonresistant Cambro-Ordovician limestones of the Cumberland Valley, and it

has been so used by most geologists working in the area.

The upper boundary of the Big Spring Station member is arbitrarily drawn at the top of the uppermost orthoquartzite or very arenaceous limestone within the lower group of Conococheague sandstones. In general the yellow-weathering silty dolomite is not found above this horizon. There are few sandy layers very high in the Conococheague of Pennsylvania and Maryland. In the excellent Waynesboro Dairy section a sandy dolomitic interval occurs 350 feet above the top of the member, and in the almost complete Big Spring Station section of the Conococheague a similar sequence occurs about 650 feet above it, but these are exceptions. From Frederick County, Virginia, northward the upper limits of the Big Spring Station member are well defined. Butts (1940) indicates that in Virginia sandstones are irregularly distributed throughout the Conococheague, but such has not proved to be the case in the area studied by the writer.

Similarly the lower limit of the member and of the Conococheague formation may be arbitrarily drawn at the lowest of these sandstone beds. The buff-yellow weathering silty dolomite in the Big Spring Station member is the same type as that in the underlying Elbrook formation, and the boundary between the two units is gradational if the sandy layers are not considered.

Errors in mapping position of Big Spring Station member.—There has been some misinterpretation of the stratigraphic position of the lower Conococheague member. Grimsley (1916), Bassler (1919), and Cloos (1951) have identified it with upper sandstones of the formation known to be present in Virginia and West Virginia. Identification of fossils in the formation is commonly necessary to distinguish the arenaceous sequences. Trilobites of Dresbachian age are found always below in the Elbrook and in limestones interbedded with the orthoquartzite beds of the Big Spring Station member. Franconian trilobites occur about 300 feet above the top of the member. On the other hand, Trempealeauian trilobites and brachiopods occur associated with the upper sandy beds of the Conococheague when the latter are present. The section at Winchester, Virginia, demonstrates the complete stratigraphic relations, for here trilobites of all three Upper Cambrian stages are present, and both sequences of orthoquartzite beds are well developed, one about 1400 feet above the other.

Grimsley's work in West Virginia (1916, p. 285) places most of the orthoquartzites of the Conococheague within the upper part. His best examples of the "upper" sandstones of the Conococheague are those on the hill south and east of Dam No. 5 on the Potomac River. Franconian trilobites (new and not yet identified) found at this section on the west limb of the anticline whose axis lies at the top of the hill indicate that the sandstones at the axis (a few hundred feet stratigraphically below the fossils) most probably belong to the Big Spring Station member. Bassler (1919, p. 80) similarly misinterpreted this member in the same fold north of the river. He considered these very arenaceous and dolomitic beds to be an anomalous occurrence of "600 feet or more of massive sandy dolomite . . . intercalated between the usual sandy laminated limestones [of the Conococheague] and the overlying Beekmantown limestone." This is the key bed considered by Cloos as high in the formation and extensively used in mapping the Conococheague in Washington County, Maryland. It is similarly well developed in the next minor fold west of Dam No. 5 and Charlton at Big Spring Station, Maryland, where even more faunal evidence shows that it is actually low in the Conococheague as originally defined by Stose.

Cloos's key bed is low in the Conococheague formation also at Clear Spring, Maryland, and in the eastern belt of outcrops at Roxbury Mills and the Antietam Battlefield sections. It seems obvious that the upper sandstones of the Conococheague in northern Virginia disappear northward along strike in West Virginia, and that the lower sandstones (Big Spring Station member) so well developed along the outcrop have been mistaken for the upper ones in Maryland. A lithofacies change in the Big Spring Station member across the Appalachian trough in an easterly direction is discussed below in a comparison of these sands with those of the Gatesburg.

Lithic details, Middle and Upper Conococheague.—The Conococheague formation except

for the basal member is a monotonous repetition of laminated limestone. Laminae of silty dolomite and siltstone from 1 mm to about 5 cm thick cause this distinctive feature. Weathering makes some of these laminae less resistant than the limestone, and commonly the rock has a banded or ribby appearance although its bedding is never platy. Given enough siliceous material within the laminae, the situation is reversed, and they form crinkly wavy ridges on the weathered surface. Bedding is generally massive. The typical fresh limestone is dark blue, weathers to a pale slate-blue or bluish-white, and is finely to medium crystalline. The lamination shows poorly on fresh exposures, but weathering quickly stains the silty dolomite layers buff. Where a bedding plane is extensively exposed along one of the coarser laminations, ripple marks, mud cracks, or burrows may be observed (Butts, 1940, Pl. 20, fig. 73).

Stose (1909, p. 6) and Bassler (1919, p. 78, 80) describe beds of "pure fine-grained rock similar to the pink limestones of the overlying formation" (Beekmantown) in the upper part of the Conococheague. The writer noted these at several sections.

Much oölitic limestone is present in the Conococheague both in upper and lower beds (Cloos, 1947). Edgewise flat pebble intraformational conglomerates are common throughout this formation, occurring with the upper pink limestone as well as toward the base. Bassler (1919, p. 86–88) describes them in some detail. The pebbles show different degrees of rounding but are generally flat and somewhat worn at the ends. They are invariably more finely crystalline than their matrix and are commonly aphanitic.

The writer has not studied the cherts of the Conococheague closely. Stose (1909) and Bassler (1919) both mention scoriaceous chert as particularly characteristic of weathering in the lower member. The writer also noted a mealy porous chert, in places resembling sandstone, associated with the weathering products of these beds.

Upper sandstones of the Conococheague and its upper boundary.—As previously stated, the upper sandstones of the Winchester, Virginia, and West Virginia sections are not abundant in the sections in Maryland and Pennsylvania. Well-exposed sections of the Conococheague above Trempealeauian trilobite genera are present at Kendall farm (47-22w), Big Spring Station (47-29w), at the type section at Scotland, Pennsylvania, at the excellent Waynescastle Dairy section (47-26w), and at Roxbury Mills. One or two thin sandstones are present high in the formation in some of these sections; none were noticed in others.

Perhaps south of northern Virginia these sandstones thicken. More paleontologic evidence is needed in southern Virginia and eastern Tennessee to demonstrate the stratigraphic position of all arenaceous beds noticed by Butts and others, but sandstones range through as much as 500 feet of the upper Conococheague in Tennessee where they have been placed in the Bloomingdale member of the Conococheague-Copper Ridge formation (Oder, 1934, p. 478) or mapped as the basal beds of the Chepultepec dolomite (Rodgers and Kent, 1948, p. 19).

In the Winchester section (45-1wa) at the southern end of the writer's area, the upper boundary of the Conococheague is placed in a covered interval of several hundred feet between U. S. Route 11 just south of the city and a sheet-metal warehouse beside a large red-brick apple-processing plant. Back of the building, typical Conococheague silt-laminated limestone crops out. These upper limestones, 240 feet thick, overlie a limestone unit of about 300 feet which includes sandstones and bears the trilobite *Prosaukia*. The limestones of the upper unit exposed behind the warehouse have yielded large unidentifiable trilobite fragments which are probably Cambrian. The lowest Ordovician fossils in this vicinity occur in beds along U. S. Route 11 just south of the railroad tracks mentioned above (*Tetralobula;* Butts, personal communication) and considerably above the upper 240-foot unit of the Conococheague. Butts mentions a Trempealeauian trilobite, *Tellerina wardi*, from the Winchester section only 150 feet (estimated by him) from the top of the formation, but the writer did not find this species in his own Winchester section.

The boundary between the Upper Cambrian Conococheague and Lower Ordovician Stone-

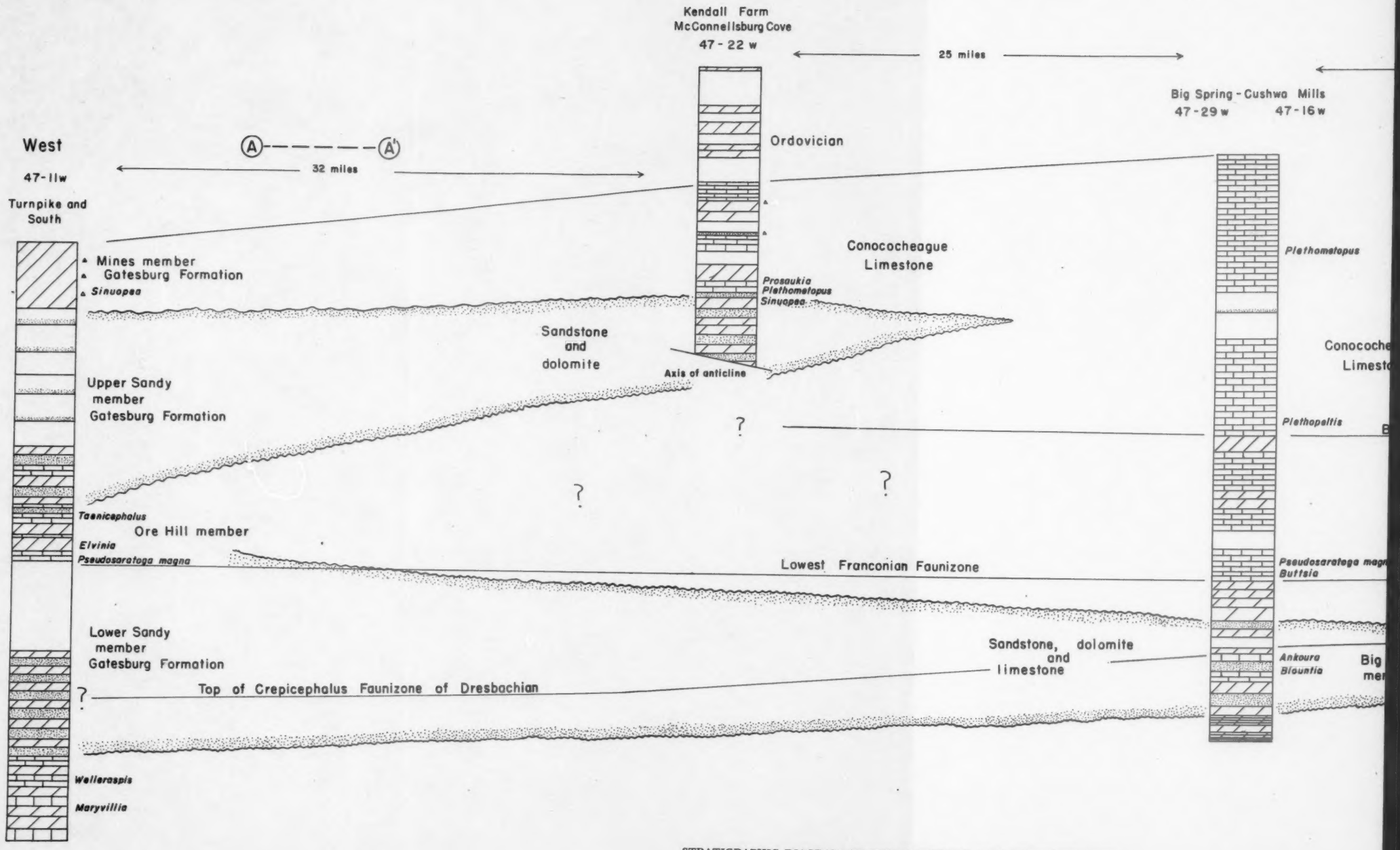
henge limestone northward in Maryland and Pennsylvania is even more difficult to draw. Bassler (1919, p. 97) gives a detailed discussion of the basal member of the Stonehenge. It consists of pure, unlaminated, massive light-colored limestone, and in places bears a typical lower Gasconade (Tribes Hill) fauna. In Maryland a middle member of the Stonehenge contains laminated limestone exactly like that of the Conococheague and may be easily mistaken for the Conococheague especially as the lowest Stonehenge beds are not so well-exposed as those above and below. The Stonehenge and its Virginia equivalent, the Chepultepec, are fossiliferous and commonly field examination of the fauna is necessary to choose an upper limit for the Conococheague which is restricted to beds of Late Cambrian age. In many sections the writer examined there were beds along the conformable boundary whose proper assignment was in doubt and probably the contact is everywhere gradational.

Relations of the Gatesburg and Conococheague orthoquartzites.—The Gatesburg formation is much more arenaceous and dolomitic than its eastern equivalent, the Conococheague, and also varies along its own outcrop; it is more calcitic and less arenaceous at the southern end of its outcrop than northward.

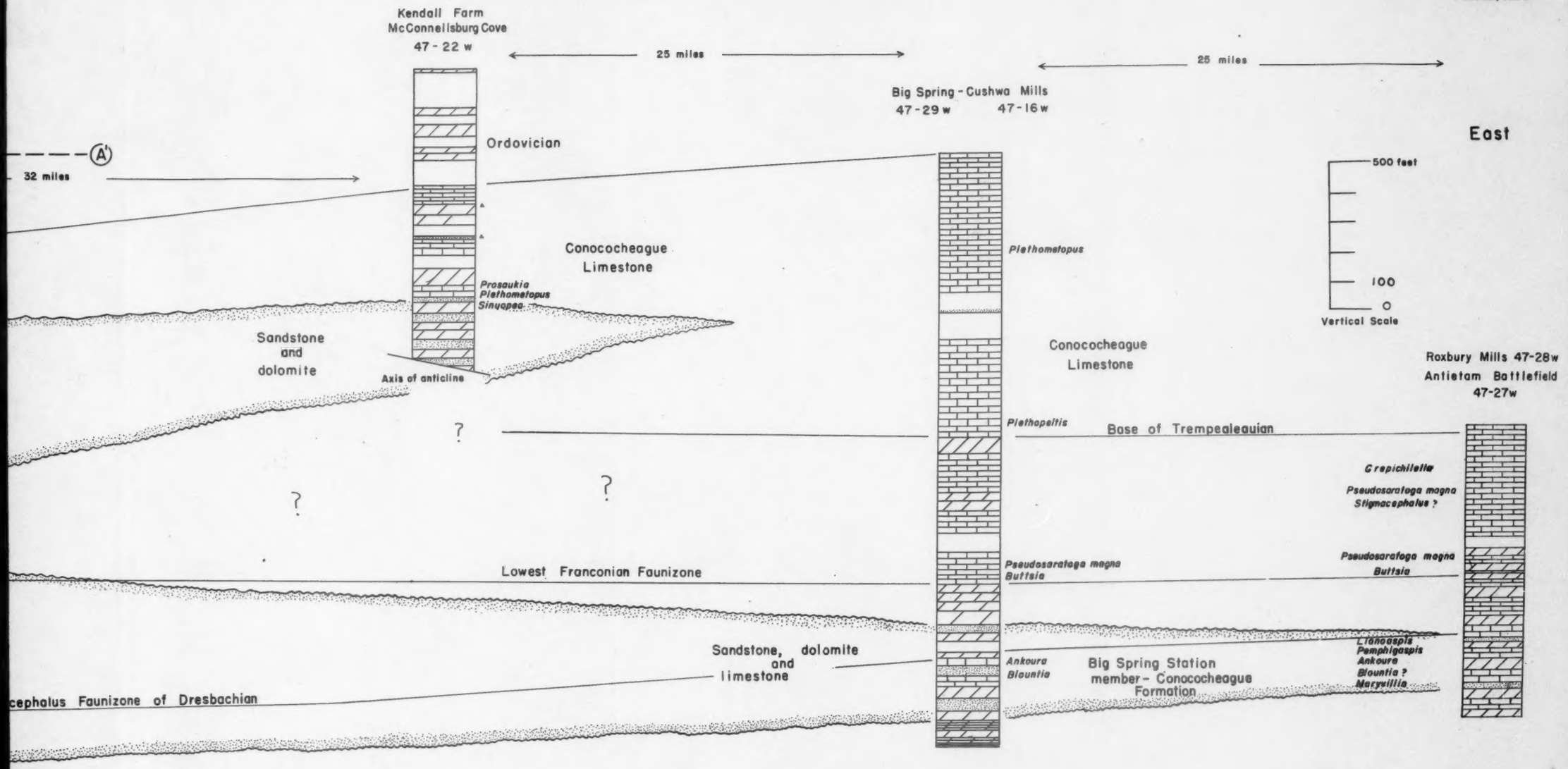
Evidence indicates that the Gatesburg rock types grade southeastward into the Conococheague-type limestone over a distance of about 40 miles. Younger sediments cover the transition beds except in a small area along the axis of a reverse-faulted anticline in McConnellsburg Cove, Fulton County, Pennsylvania. This area lies roughly east of the southernmost measured Gatesburg section (on Pennsylvania Turnpike). The Kendall Farm section (47-22w), located on a cross road following Kendall Run between the Cito road and U. S. Route 522 about 3 miles south of McConnellsburg, is doubly exposed on both limbs of the anticline; dips of about 60° SE. prevail along the axis which crosses the creek slightly northeast of the farm house. The west limb of the fold shows almost 400 feet of partly covered cherty light and dark crystalline dolomite (Ordovician ?) overlying 370 feet of typically laminated Conococheague-type limestone, with abundant cryptozoon stromatolites and several 50-foot intervals of dolomite with but little sand. The base of this interval contains the Trempealeauian genera *Prosaukia*, *Sinuopea*, and *Plethometopus* (?). Below this section and occupying a large area along the fold axis is a very poorly exposed interval probably of orthoquartzite and dolomite deeply weathered into a Gatesburg-type ridge covered with sandstone float, trees, brambles, etc. This may represent at least 200 feet of the upper Gatesburg sandy member. There appears to be more sandstone here than in the arenaceous beds of the typical Conococheague farther east. These beds do not support such a densely wooded ridge with such sandy soil. The Conococheague limestone apparently replaces the Gatesburg lithofacies eastward (Pl. 3).

Gatesburg and Conococheague orthoquartzites are closely similar in grain size. Thirty Conocheague specimens examined (mostly from the Big Spring Station member) and 13 Gatesburg (lower sandy member) specimens show almost identical histograms. A typical analysis shows about 7 per cent coarse, 45–50 per cent medium, 25–30 per cent fine, and 10–15 per cent very fine quartz sand. The textures of upper Conococheague sands differ little from those of the lower member. Prouty (1948a, p. 1603; 1948b, p. 1344) compared sandstones from the Copper Ridge (Gatesburg equivalent of Tennessee) with those of the Conococheague and found a northwest coarsening, but the writer's specimens from the central Appalachians do not show this. However, sandstones in the Gatesburg and Conococheague probably had a common origin, and similar processes must have produced and transported them. Most of Prouty's generalizations on the Copper Ridge sandstones of Tennessee (1948a, p. 1603) hold for those of the central Appalachian equivalents. Both Gatesburg and Conococheague sand grains are well rounded and frosted.

The Gatesburg formation becomes more sandy northwestward, and the apparent southeasterly decrease in sand content of individual arenaceous beds seems to continue in sections of the Conococheague. Data (based on unweathered specimens) show that the Gatesburg orthoquartzites at Flucht-Eschelmann section (47-8w) are about 55 per cent quartz sand. Sandstone beds in the westernmost Conoco-



STRATIGRAPHIC DIAGRAM OF THE GATESBURG AND CONOCOCHEAGUE FORMATIONS



STRATIGRAPHIC DIAGRAM OF THE GATESBURG AND CONOCOCHEAGUE FORMATIONS

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cheague section examined (Big Spring Station, 47-29w) and three sections in the eastern Conococheague belt of the Valley show a progressive decrease in sand content southeasterly. Sections in southern Pennsylvania show about 36 per cent quartz sand in lower Conococheague beds, and the Roxbury Mills section south in Maryland only 33 per cent.

The thickness and general rock content of the Big Spring Station member also change eastward across the Shenandoah-Cumberland Valley. The most westerly sections show the thickest, purest, and most numerous orthoquartzite beds. The Winchester (45-1w), Big Spring Station (47-29w), and Musser Farm (47-23w) sections expose numerous quartzite beds up to 5 feet thick, but sections at Roxbury Mills (47-28w) and Antietam Battlefield (47-26w) in the southeasternmost area examined contain no true orthoquartzite beds but only a few scattered layers of finely arenaceous siliceous dolomite which form a topographic ridge at the base of the formation just as do the more arenaceous beds to the north and southwest. Sandy beds at the type section occur through 310 feet, those at Winchester, Virginia, through about 370 feet. None of the other sections are as well exposed nor so nearly complete, but arenaceous or siliceous dolomite beds in them range through from 100 to 240 feet.

Sections of the Conococheague formation.—Locality 47-22w, Kendall Farm in McConnellsburg Cove: On a cross road following Kendall Run between the Cito road and U. S. Route 522 about 3 miles south of McConnellsburg, Fulton County, Pennsylvania. The section was measured on slopes above the creek on the north side of the valley behind two farm houses. It is an important locality because it shows both the Conococheague and Gatesburg facies. It is exposed on both limbs of an anticline which exposes the Gatesburg facies along its axis.

PART 1: Conococheague limestone on east limb of anticline; dips 30°–40° S. 65° E. proceeding toward fold axis

	Feet
White dolomite and a little limestone, medium-bedded and coarsely sandy	3.0
Covered interval	86.0
Massive cryptozoon limestone, 47-22w.3	2.0
Platy limestone	5.0
Covered interval	22.0
Coarsely sandy bluish-weathering limestone	11.0
Bluish platy limestone, typical Conococheague with edgewise conglomerate and blobs of black chert	8.0
Oölitic chert	.2
Highly cherty limestone with edgewise conglomerate	2.0
Cryptozoon limestone	1.0
Covered interval	2.0
Platy bluish limestone, edgewise conglomerate and cryptozoon	10.5
Pure massive-bedded light-colored dolomite	16.0
Bluish limestone	3.0
Dolomite	1.0
Covered interval	42.0
White dolomite	21.0
Covered interval	60.0
47-22w.2, coquina of *Prosaukia* Bluish-weathering limestone, oölitic and fossiliferous	39.0
47-22w.1, *Plethometopus* and associated pygidium of same type as found high at Big Spring Station; also gastropods.	
Total Conococheague facies	335.0

The next 1100 feet west along the ground is covered with sandstone float, brambles, and trees and forms a topographic ridge like that of the typical Gatesburg farther west in the Everett quadrangle. It is estimated that from 200 to 400 feet of the upper sandy member of the Gatesburg supports this ridge. The fold axis lies behind and slightly northeast of the Kendall farm house.

PART 2: Cambro-Ordovician formations on west limb of the anticline; dips from 60° N. 72° W. at axis to 68° N. 65° W. at far west flank

Ordovician dolomites:

	Feet
Pure white dolomite, coarsely sandy at base and cherty at top	18.5
Covered interval	106.0
Partly covered, but probably all light and dark crystalline dolomite	163.0
Covered interval	111.0
Total Ordovician dolomite	398.5

Conococheague limestone:

Typically laminated Conococheague limestone with edgewise intraformational conglomerate and unidentifiable trilobite fragments	46.0
Typical Conococheague limestone, irregularly laminated	30.0
Mostly dolomite with black chert nodules	59.0
Covered interval	23.0
Cryptozoon limestone and some dolomite with a 3-inch bed of sandstone and some stringers of chert and sandstone in the limestone	38.0
Alternating limestone and dolomite with chert stringers	37.5
Cryptozoon limestone	3.5
Covered interval	36.5
Mostly dolomite	62.0

Blue-weathering fossiliferous limestone..... 47-22w.1a, *Prosaukia, Sinuopea*	32.0
Total Conococheague above Trempealeauian fossils	367.5

LOCALITY 47-25w, South of Waynesboro, Pennsylvania: 2 miles south of town on Hagerstown road where it crosses Antietam Creek at the junction of its two tributaries. The Elbrook formation is seen in exposures from the cemetery in the south outskirts of Waynesboro along the road to the bridge across the creek. The base of the Big Spring Station member is close to the north end of the bridge where weathered arenaceous beds show the lowest sandstones in the sequence below Dresbachian trilobites. The position in the section of the fossil locality in the Elbrook 700–800 feet below the base of the Big Spring Station member is estimated from striding the distance in direction of dip and applying an average of the dips measured in the road cuts.

From the south end of the bridge upward the Big Spring Station member is exposed bed-by-bed in the deep road cut (Dip 21° S., 18° W.)

Conococheague limestone

	Feet
66. Limestone (top exposure in cut).....	1.0
65. Covered intervals.....	2.0
64. Limestone.....	2.0
63. Covered, but likely shaly dolomite.....	6.0
62. Very finely laminated dolomite.....	3.0
61. Massive finely crystalline limestone.....	2.5
60. Massive dolomite.....	1.0
59. Platy, shaly dolomite.....	4.0
Conococheague limestone above Big Spring Station member.....	21.5

Big Spring Station member

58. Coarse sandstone.....	.5
57. Laminated limestone with chert in top..	4.5
56. Platy, shaly dolomite.....	8.5
55. Massive dolomite.....	1.0
54. Laminated limestone.....	3.0
53. Covered interval.....	1.0
52. Laminated limestone with shaly dolomite intervals.....	2.0
51. Laminated limestone.....	5.0
50. Laminated limestone.....	5.5
49. Dolomite.....	1.0
48. Laminated limestone.....	4.0
47. Coarsely oölitic limestone.....	2.0
46. Sandstone grading up into sandy dolomite.....	1.0
45. Massive crystalline limestone.....	5.0
44. Massive dolomite.....	1.0
43. Shaly dolomite.....	.5
42. Massive limestone.....	6.0
41. Finely laminated dolomite.....	2.0
40. Coarse sandstone with dolomitic matrix	1.0
39. Coarsely sandy dolomite.....	2.0
38. Finely laminated dolomite.....	4.0
37. Limestone.....	.5
36. Coarsely sandy laminated dolomite.....	3.0
35. Massive laminated limestone.....	4.0
34. Finely laminated dolomite.....	.5
33. Limestone.....	1.0
32. Thin-bedded laminated dolomite.....	2.5
31. Massive crystalline limestone..... 47-25.4 *Ankoura, Tricrepicephalus* (?)	3.5
Big Spring Station member above top Dresbachian fauna.....	75.5
30. Very finely laminated dolomite.....	6.5
29. Laminated limestone.....	3.0
28. Massive bedded finely crystalline dolomite.....	2.0
27. Shaly dolomite with some limestone....	6.0
26. Dolomite impregnated with calcite.....	.5
25. Laminated limestone.....	5.0
24. Massive coarsely crystalline limestone..	3.5
23. Laminated limestone and dolomite.....	2.0
22. Massive banded dolomite with some limestone (beds disrupted by blasting)	5.5
21. Laminated limestone and dolomite..... 47-25w.3, *Kingstonia* and *Tricrepicephalus*.	3.0
20. Massive laminated limestone.....	2.5
19. Medium-bedded coarsely crystalline oölitic limestone.....	1.5
18. Laminated dolomite, some limestone....	1.5
17. Laminated limestone.....	1.0
16. Massive finely crystalline dolomite.....	2.0
15. Platy, shaly dolomite.....	1.0
14. Laminated limestone.....	7.0
13. Covered interval across creek.....	95.0
12. Dolomite with limestone beds at top....	5.0
11. Covered with sandstone float.....	10.0
10. Poorly exposed sandstone, probably weathered sandy dolomite.....	1.0
Big Spring Station member with Dresbachian fauna.....	164.5
Total Big Spring Station member.....	240.0

Elbrook formation:

9. Massive and platy dolomite.....	5.0
8. Limestone.....	1.0
7. Massive dolomite.....	4.5
6. Covered interval.....	6.0
5. Medium-bedded fine-grained dolomite..	2.0
4. Covered interval.....	3.0
3. Platy dolomite.....	.5
2. Massive limestone.....	.5
1. Massive dolomite.....	4.0
Partly exposed Elbrook.....about 47-25w.2, *Wellerasps* (?)	400.0
Partly exposed Elbrook (road cut by Harrison School).....about 47–25w.1, *Maryvillia* (?), *Alokistocare, Blountia, Ankoura* in laminated limestone.,	350.0
Total Elbrook approximately.....	800.0

LOCALITY 47-26w, Waynescastle Dairy: 3½ miles east of Greencastle at the railroad crossing of Pennsylvania State Highway 16 between Greencastle and Waynesboro, Pennsylvania, southern Chambersburg quadrangle. The top of the section is essentially at the top of the Conococheague. The highway underpass of the railroad separates the Conococheague exposures from those of the Stonehenge lying west of the tracks. The Conococheague is exposed in vertical beds in the deep cuts along the highway east of the railroad for about one half mile. The direction of dip ranges from N. 85° W. to N. 80° E. The lowermost exposures of the Conococheague dip 52°–44°, but proceeding up in the section (west) the dip soon changes to vertical. The lowest Conococheague is incomplete and poorly exposed, but may be examined in the road cuts 4.5 miles away, just south of Waynesboro (47-25w). The Waynescastle Dairy is the best exposure of the Conococheague formation in Pennsylvania. It is particularly important since the type section is structurally deformed and incomplete (Cloos, 1951, p. 46). Much of this section may be measured bed-by-bed and, although it is not quite complete, one may get a good idea of lithofacies and thickness. Because the beds are relatively unweathered and vertical no fossils could be obtained.

The top of the Conococheague limestone:

	Feet
144. Thin-bedded laminated limestone	90.0
143. Coarsely crystalline limestone	8.0
142. Massive dolomite	10.0
141. Covered interval	7.0
140. Laminated limestone	21.0
139. Finely laminated dolomite and siltstone	3.0
138. Shaly dolomite	1.0
137. Black chert and limestone	.5
136. Thin-bedded laminated limestone	25.0
135. Coarsely crystalline limestone	1.5
134. Thin-bedded laminated limestone	6.0
133. Coarsely crystalline limestone	1.5
132. Thin-bedded laminated limestone	4.0
131. Coveral interval, shale or siltstone	.5
130. Coarsely crystalline, laminated oölitic limestone	6.0
129. Coarsely crystalline limestone, few laminae, intraformational edgewise conglomerate	6.0
128. Finely laminated dolomite	2.0
127. Laminated limestone, crystalline and rather pure at base; top with more dolomitic silty laminae	38.5
126. Coarsely crystalline limestone with oölite and edgewise conglomerate	4.0
125. Finely laminated dolomite	1.0
124. Laminated limestone	3.5
123. Oölitic limestone	6.0
122. Coarsely laminated crystalline limestone	15.0
121. Coarsely crystalline limestone	2.0
120. Oölitic limestone	.5
Conococheague limestone above sandstone	263.5
119. Dark calcareous coarse sandstone	2.0
118. Coarsely sandy dolomite	2.0
117. Pure limestone with little dolomite lamination	19.5
116. Finely laminated dolomitic limestone, some pure dolomite	10.5
115. Edgewise intraformational conglomerate and oölitic limestone	2.5
114. Edgewise intraformational limestone conglomerate	1.0
113. Laminated limestone	4.0
112. Massive crystalline limestone	5.5
111. Finely laminated limestone	5.5
110. Massive unlaminated crystalline limestone	8.0
109. Laminated dolomite	.5
108. Limestone	.5
107. Laminated dolomite	1.0
106. Laminated limestone	16.0
105. Oölitic limestone	2.0
104. Edgewise intraformational conglomerate	.5
103. Medium-bedded dolomite	7.0
102. Laminated limestone	29.0
101. Covered interval but float shows silty dolomite	25.0
100. Laminated limestone	25.0
99. Massive edgewise intraformational limestone conglomerate	2.0
98. Massive dolomite	2.0
97. Finely laminated limestone	24.0
96. Massive unlaminated limestone	2.5
95. Laminated limestone	40.0
94. Massive crystalline limestone with intraformational conglomerate. Large 'B' painted on rock	4.5
93. Laminated limestone (at east edge of lower cut and west end of bridge)	23.0
Typical Conococheague limestone	265.0
92. Covered interval where creek crosses highway	386.0
Beds described below are in meadow just north of road and at east side of creek flat. Section soon reverts to highway cut and is measured on north side.	
91. Laminated limestone	25.0
90. Covered interval	20.0
89. Covered interval, some laminated limestone	98.0
88. Partly covered interval (located in field to north of upper cut and 75 feet east of its west end), laminated limestone	25.0
87. Finely laminated dolomite (located in road cut)	4.0

86.	Typically laminated Conococheague limestone	36.5
85.	Alternating laminated limestone and dolomite	12.0
84.	Massive limestone	3.5
83.	Dolomite	.5
82.	Coarsely laminated crystalline limestone	6.0
81.	Laminated dolomite	5.5
80.	Laminated limestone	5.5
79.	Massive unlaminated limestone with edgewise intraformational conglomerate	2.5
78.	Finely laminated limestone	8.0
77.	Light-colored, brown-weathering pure dolomite	3.0
76.	Finely laminated limestone, typical Conococheague (500 laminae in 3 feet)	41.0
75.	Coarsely laminated limestone, typical Conococheague, some crystalline limestone	13.0
74.	Typical Conococheague limestone with more silty laminae than is usual. Lower part of interval covered. Mainly buff silty laminated dolomite which weathers to platy, shaly, reddish material	100.0
73.	Dolomitic siltstone	2.5
72.	Typical Conococheague with more buff silty laminae than usual	24.0
71.	Buff silty, finely laminated dolomite and limestone	3.5
70.	Typical Conococheague limestone with fine laminae	17.0
69.	Platy buff finely laminated dolomite	1.0
68.	Laminated limestone with cryptozoon	3.0
67.	Finely laminated typical Conococheague limestone	8.5
66.	Buff-weathering finely laminated dolomite	3.0
65.	Laminated Conococheague type limestone	12.5
64.	Silty and finely sandy dolomite	.4
63.	Finely crystalline buff dolomite	.6
62.	Laminated limestone	9.0
61.	Finely laminated buff dolomite	1.0
60.	Laminated limestone	7.5
59.	Buff laminated dolomite	1.5
58.	Coarsely sandy buff dolomite, weathers to porous sandstone	1.5
57.	Typical laminated Conococheague limestone. Coarse laminae of silty dolomite which weather brown while the limestone weathers blue, giving exposed surface a banded appearance. Laminae range from several millimeters to 3 or 4 cm thick and are irregular. Limestone intervals are a little thicker than silty layers. On weathered outcrop here the limestone layers stand out	25.0
56.	Laminated dolomite	1.0
55.	Irregularly laminated limestone	12.0
54.	Laminated calcareous dolomite	1.0
53.	Coarsely sandy buff laminated dolomite	2.5
52.	Laminated limestone (48 laminae in 3 feet)	3.0
51.	Brown-weathering laminated dolomite	1.5
50.	Massive unlaminated limestone	10.5
49.	Laminated dolomite	.5
48.	Finely laminated limestone	1.5
47.	Buff massive dolomite	1.5
46.	Very finely laminated limestone with cryptozoon (40 laminae to an inch)	3.0
45.	Massive limestone, irregular laminae not prominent	4.5
44.	Covered interval	1.5
43.	Finely laminated dark silty dolomite (2 laminae to an inch)	2.0
42.	Covered interval	1.0
41.	Massive limestone	2.0
40.	Covered interval	83.0
39.	Massive limestone and dolomite	17.0
38.	Massive limestone with cryptozoon at top	5.0
37.	Mostly covered but likely massive limestone	7.0
36.	Calcareous dolomite	2.0
35.	Massive dolomite	4.0
34.	Platy, shaly buff-colored dolomite	5.0
33.	Covered interval	6.0
32.	Calcareous massive dolomite	2.0
31.	Medium-bedded to platy thin-bedded dolomite	21.0
30.	Platy-weathering silty dolomite	8.0
29.	Massive dolomitic limestone	10.0
28.	Covered interval	63.0
27.	Massive and platy buff dolomite; dark when fresh, weathers brown	6.0
26.	Dolomitic massive limestone	5.0
25.	Covered interval	2.0
24.	Massive dolomitic limestone	3.0
23.	Laminated limestone	3.5
22.	Silty dolomitic limestone	2.0
21.	Platy, shaly dolomite	1.0
20.	Massive dark aphanitic limestone	4.5
19.	Dark aphanitic limestone	2.5
18.	Buff-weathering dolomite	4.5
17.	Coarsely crystalline limestone	1.5
16.	Buff-weathering dolomitic limestone	1.0
15.	Medium-bedded silty dolomite, upon weathering becomes chippy and buff-colored	7.0
14.	Pure finely crystalline limestone	2.0
13.	Chippy and platy-weathering silty dolomite, brownish	5.5
	Typical Conococheague limestone	863.5

Big Spring Station member of the Conococheague limestone:

12.	Calcareous sandstone	1.5
11.	Massive limestone	5.0
10.	Laminated limestone, usual Conococheague type (these beds exposed in field north of east end of road cut)	50.0
9.	Covered interval	57.0
8.	Sandstone indicated by float	2.0
7.	Massive limestone with intraformational conglomerate (this is last exposure at east end of cut)	4.0
	Total Big Spring Station member	119.5

The Big Spring Station member may be incom-

pletely exposed. East of this point an anticlinal axis repeats the section.

Total Conococheague formation exposed..... 1897.5

Faunas and age of the Conocheague formation.—The Conococheague is not so fossiliferous as the Warrior formation and Ore Hill member of the Gatesburg. Fossils have not been found where the beds are strongly folded. In good exposures with low dips, protracted search may reveal a few inches of fossiliferous rock concealed within hundreds of feet of dark limestone much like that of the Ore Hill but more massive. Generally the fossiliferous limestone is more crystalline than other beds and commonly weathers with a minutely etched surface. Even fossils collected at the best localities are distorted to some degree in accordance with the structure of the beds. Proportions are altered, and angles changed, and in some instances distortion is so great that forms are bent back upon themselves or stretched. In all instances plastic movement of the rock has partially destroyed the texture of the replaced test so that fine ornamentation is largely obliterated and an almost fibrous texture appears under microscopic examination.

Despite the above difficulties, there is ample evidence that the Conococheague as defined by Stose almost spans the complete Upper Cambrian sequence. The Big Spring Station member represents the *Crepicephalus* faunizone (Middle Dresbachian) already discussed with the Elbrook faunas. No *Aphelaspis* zone faunas have been found, but from 50 to 500 feet above the top of the Big Spring Station member occurs the lower Franconian *Pseudosaratogia magna* faunizone. These species have already been described (Wilson, 1952) and the stratigraphic relationships with the Ore Hill member made known. The fauna commonly consists of:

Bullsia drabensis Wilson†
Drabia curtoccipita Wilson*
Drabia sp. undet.
Dellea saratogoensis (Resser)†
Eshelmania snoburgensis Wilson*
Pseudosaratogia magna Wilson*
Cheilocephalus sp. undet. (?)†

* Species in common with the *Pseudosaratogia magna* fauna at the base of the Ore Hill.

† Species in common with the typical *Elvinia* fauna of the Ore Hill member.

The writer believes (Wilson, 1951) that the *Pseudosaratogia magna* faunizone represents the entire lower Franconian and possibly also middle Franconian time in the eastern part of the Appalachian geosyncline. Here it is thus equivalent at least to the *Elvinia* zone, which occurs above it in the Ore Hill member. Probably, as previously outlined the faunas represent two biofacies expressions of the same time-stratigraphic unit. Trilobites of the *Pseudosaratogia magna* faunizone are found consistently in the lower third of the Conococheague through as much as 130 feet of section. A new species, *Stigmacephalus ? distorta*, occurs high in the zone with *P. magna* at Roxbury Mills and Antietam Battlefield sections (in the latter section misprinted "*Crepichilella antietamensis*" at 47-27w.2 in Wilson, 1951, p. 623). Shortly above it at Antietam Battlefield section (47-27w.3) the species *Crepichilella antietamensis* occurs alone. This may well mark a higher faunal unit. This species and *Stigmacephalus ? distorta* are probably middle Franconian trilobites.

The discovery of saukinid trilobites near Natural Bridge, Virginia (*Tellerina wardi* (Walcott) Resser, 1938, p. 101), and in Stose's original section of the formation (*Saukia stosei* Walcott, 1914, p. 383) confirmed the Trempealeauian age of at least part of the Conococheague limestone many years ago. Collections made by the writer show that the Trempealeauian stage is represented in the formation from Frederick County, Virginia, north by trilobites occurring 800 to 1400 feet above the base of the Big Spring Station member (base of the formation) and probably not much above the base of the upper Conococheague sandstones when these are present (Table 5).

Prosaukia is the only saukinid collected by the writer in the Conococheague. It is invariably associated with Trempealeauian genera and has been commonly found toward the top of the Conococheague far above established Franconian trilobites. The fauna of the upper Conococheague formation cannot be assigned to any particular Trempealeauian trilobite zones listed by Howell *et al.* (1944), partly because the standard faunal sequence of the type Trempealeauian has not been satisfactorily established. Certain "zones" of the

Croixan area appear to be facies equivalents of others (Bridge, 1937a, p. 236), and the questionable subdivision of the Saukidae by Ulrich and Resser (1933) has obscured the stratigraphic relationship of some of these

ranging and of little value. A form very close to *Cryptozoon proliferum* Hall (supposedly a lower Conococheague form) occurs as well higher in the formation and also common through the entire Gatesburg. *C. undulatum*

TABLE 5.—OCCURRENCES OF TREMPEALEAUIAN TRILOBITES IN THE CONOCOCHEAGUE FORMATION

Location	General	Stratigraphic position above base of formation
Scotland, Penna.	*Prosaukia stosei* (Walcott) *Tellerina scotlandensis* Resser	Unplaced but quite high
Kendall Farm 47-22w.1 and .2	*Prosaukia* sp. *Plethometopus* (very large sp.)	Unplaced but quite high
Winchester, Va. 45-1wa.1	*Prosaukia*	1400 feet
Big Spring Station 47-29w.3	*Plethometopus* (very large species listed above)	1300 feet
Big Spring Station 47-29w.2	*Plethopeltis* (?)	800 feet
One mile northeast of Stonehenge, Penna., 47-9w	*Prosaukia*, *Plethometopus*, *Stenopilus*, *Leiocoryphe*, *Bowmania*, orthoid brachiopods	Unplaced

genera. Further, collections from the Conococheague lack certain key genera of the Croixan region (*Dikelocephalus*, *Calvinella*, *Saukia*) and contain mostly genera ranging through the highest Franconian and Trempealeauian of Wisconsin and Minnesota.

Use of cryptozoon stromatolites as index fossils.—Some writers have considered algal stromatolites in the Appalachian Upper Cambrian as reliable index fossils. Bassler (1919) describes and photographs *Cryptozoon undulatum* and *C. proliferum* Hall favoring their use as indices to lower Conococheague time. Fenton and Fenton (1937, p. 435–441) have described several additional types. An attempt to use these algal precipitates stratigraphicaly has been made in the Middle and Upper Cambrian beds of the Lehigh Valley, northeastern Pennsylvania (Howell *et al.*, 1950).

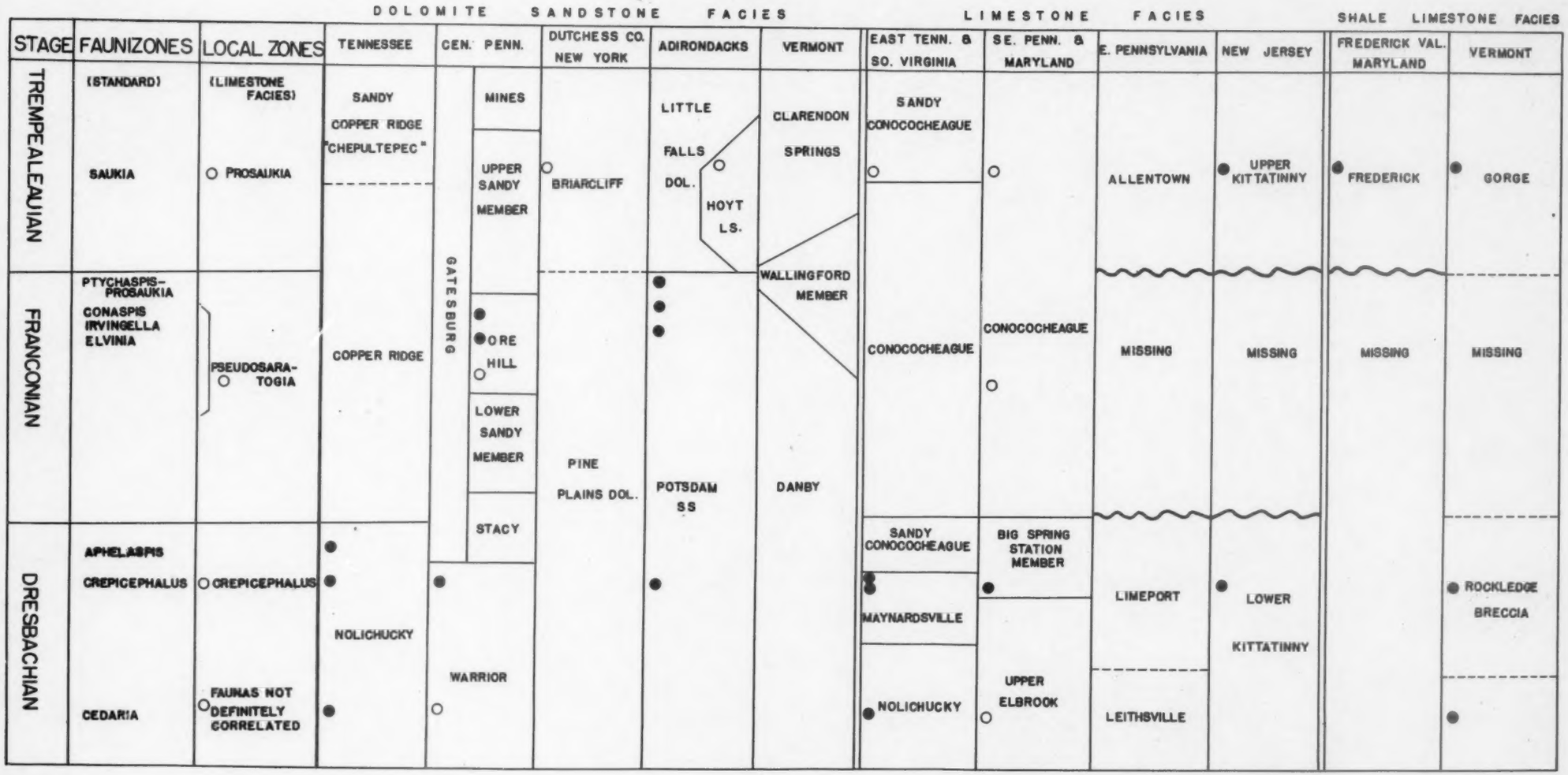
The writer has given little attention to Conococheague and Gatesburg stromatolites. Nevertheless, many types abound in these formations, and there is good opportunity to check their usefulness as index fossils as compared to trilobites. The algae are obviously environmentally controlled; in the Gatesburg they are confined to the uppermost (purer) dolomite lithotope of the cycle, and in the Conococheague to the less clastic limestones. Evidence exists that some species may be long

(*Archaezoon undulatum* (Bassler)) is found with Dresbachian trilobites in the Lehigh Valley Limesport formation (Howell *et al.*, 1950, p. 1395) and in the Dresbachian Warrior formation of the Nittany Arch. It is present in the lower Conococheague with Dresbachian and Franconian trilobites (Cushwa Mills section, 47-16w) and has been found at the Kendall Farm section (47-22w) above Trempealeauian trilobites near the top of the formation. Additional work is needed in the central Appalachians to distinguish the several types of stromatolites outlined by Howell *et al.* (1950, p. 1359) and to check their validity as index fossils.

FORMATIONS OF THE FREDERICK VALLEY

Frederick Limestone

Bassler (1919, p. 115) gave the name Frederick limestone to the platy-weathering "building or fence stone" common around Frederick, Maryland. The wide valley crossing Frederick County, Maryland, north of the Potomac River is downfaulted from the South Mountain-Catoctin Mountain anticlinorium on the west and rimmed to the east by the Martic overthrust, the thrust plate of which is formed largely of Precambrian or early Paleozoic sediments. The Frederick and overlying Lower



CORRELATION TABLE, UPPER CAMBRIAN IN THE CENTRAL AND NORTHERN APPALACHIANS

Ordovician Grove limestone are the eastern-most unmetamorphosed Cambro-Ordovician beds of the central Appalachians and are separated from the Lower Paleozoic exposures in the Cumberland-Shenandoah Valley by about 12 miles across the overthrust and highly folded South Mountain anticlinorium. The Frederick Valley limestone may be traced northward into Hanover Valley, Pennsylvania, to form part of the similar Conestoga formation (personal communication by Stose and Stose). Stose and Stose (1946) have greatly refined and supplemented Bassler's early work.

Stose and Stose (1946, p. 43, 44) describe the Frederick formation as "thin-bedded blue limestone, slabby granular limestone with many argillaceous partings, buff earthy limestone, and blue mottled argillaceous limestone" with some "thin dark shale in the lower part of the formation." They state that it unconformably overlies the Lower Antietam quartzite easterly in the valley. The formation is estimated from numerous incomplete and poorly exposed outcrops to be about 500 feet thick.

The writer has examined the Frederick limestone only in the eastern environs of the city of Frederick; fossils were collected from slabs in the stone fence by the meat-packing plant near the top of the formation. The impure thin-bedded limestone here yields about 7 per cent very fine sand and 3 per cent silt by volume. A large amount of organic matter is present and a black scum results from acid treatment.

Although Bassler (1919, p. 117) considered the Frederick as Chazyan or early Mohawkian, later work has indicated a Late Cambrian age (Stose and Stose, 1946, p. 46). The brachiopod "*Strophomena*" *stosei* Bassler proved to be a new genus of early orthoid, *Xenorthis*, when studied by Ulrich and Cooper (1938, p. 31, 74, 75). The following trilobites have also been identified from poorly silicified material collected by the writer, Dr. William Evitt, and Dr. Franco Rasetti: *Plethometopus*, *Stenopilus*, *Liptychaspis*, and the spinose free cheek "*Acidaspis ulrichi*" of Bassler. All these indicate a Trempealeauian age; "*A. ulrichi*" has been found also in the Gorge formation by Rasetti and in the highest Cambrian of Utah by G. A. Cooper and Hintze. A Late Cambrian age of the Frederick is more in keeping with the areal geology in the east part of the city. Here the slabby fossiliferous limestone appears to dip conformably beneath the Grove formation which bears Lower Ordovician fossils at several places.

Regional Relations of the Upper Cambrian Formations of the Appalachian Mountains

Introduction

The isopachous facies map (Fig. 4) is based on the correlation chart of Appalachian Upper Cambrian formations (Pl. 4). Franconian and Trempealeauian formations from the three areas discussed in this paper occur in three lithofacies. A broad area of dolomite and orthoquartzite is known from outcrops of the Gatesburg formation of central Pennsylvania, from outcrops of the Potsdam sandstone around the Adirondacks, from wells in western New York and over the Findlay Arch in Ohio, and from outcrops of the Copper Ridge formation in southern Virginia and Tennessee.

A second facies, represented by a narrow belt of dark laminated limestone, extends from Tennessee to New Jersey (Conococheague, Allentown, and Kittatinny formations) and roughly follows the line of maximum thickness (2000 feet) along the geosynclinal axis. This limestone contains the eastward feather edges of certain sand bodies of the dolomite-orthoquartzite facies. Eastward, the first two facies grade into one another. The Copper Ridge–Conococheague relations have been demonstrated by Butts (1933, p. 9) and Oder (1934, p. 476), and the writer indicates that similar relations hold for the Gatesburg and Conococheague.

The silt lamination of the limestone belt increases eastward, and the third facies represented by the Frederick (and possibly also the Conestoga) limestone is predominantly very thin-bedded dark limestone with many silt interbeds. The writer believes this central Appalachian facies may be homologous with part of the slate belt Cambrian of Vermont (Rockledge and Gorge formations) now thrust

over a carbonate sequence along the east shore of Lake Champlain. Cady (1945, p. 537, 538) has demonstrated that here the sandstone and dolomite of the Danby and Clarendon Springs formations pass gradually to the northeast through the Wallingford sandy dolomitic limestone member into the "slate succession." Much the same picture is envisioned for the relations of the Frederick and Conococheague formations.

Source of Sediments in the Geosyncline

Recently, it has been realized that Upper Cambrian sediments in the Appalachian trough were probably derived from the west and northwest as well as from the east. The dolomite-orthoquartzite facies becomes sandier in the direction of the Canadian Shield (Krynine, 1948a, p. 5; Swartz, 1948a, p. 7; 1948b, p. 1050; Prouty, 1948b, p. 51). The writer's own limited study of Gatesburg and Conococheague sands showed no northeastward coarsening but indicated that sands in the two formations probably had a common source and that beds were relatively more sandy toward the shield area rather than easterly in the geosyncline. Thiel (1935, p. 559–614) has indicated that sands of the Ordovician St. Peter formation do not coarsen toward their source (also the Canadian Shield), probably because they are not first-cycle deposits but were derived from older sandstones. The Gatesburg sands resemble those of the St. Peter in degree of sorting, medium grain size, degree of rounding and frosting, and heavy-mineral suite, and are perhaps also secondary deposits derived from Late Precambrian sandstones.

If, on the other hand, the Upper Cambrian sandstones are first-cycle deposits derived from weathered igneous rocks of the shield, one must explain the paucity of finer clastics in the sequence. Clays and silts may have been removed during long-enduring wave action along dune-covered beaches (Prouty, 1948b, p. 51) or by intense wind deflation on deeply weathered Cambrian soils unprotected by a vegetative cover (Swartz, 1948b, p. 1550; Krynine, 1946, p. 15). A small part of the silt and clay may be represented by the Frederick limestone on the eastern side of the geosyncline. This formation contains more fine clastic material than any other Upper Cambrian unit, and assumedly represents a very silty aspect of the Conococheague, the postulated transitional beds having been eroded away across South Mountain anticlinorium. The silt content of the Frederick may have been derived from the eastern border land, Appalachia. The writer speculates that the easterly increasing amounts of silt and clay, whatever their source, may be the environmental factor which

LIST OF REFERENCES FOR SECTIONS SHOWN ON ISOPACHOUS-FACIES MAP (FIG. 4)

1. California Co., No. 1, A. R. Spears, Lincoln Co., Kentucky. Base and top picked by Mrs. L. B. Freeman on sample log showing insoluble residues.
2. Thorn Hill section, Thorn Hill, Tenn., along Copper Ridge, Grainger Co., (Hall and Amick, 1934; Rodgers and Kent, 1948, includes basal Chepultepec sand of these writers).
3. Jockey Creek section, Limestone, Tenn., 20 miles west of Johnson City (Oder, 1934, p. 469, includes Bloomingdale sandy member).
4. Widner Branch, 13 miles east of Abingdon, Washington Co., Va. (Butts, 1940, p. 88).
5. Buchanan Branch, southwest of Saltville, Va. (Butts, 1940, p. 93).
6. A mile southeast of Staunton, Augusta Co., Va. (Edmundson, 1945, p. 115).
7. Winchester, Va., southwest outskirts of town, Frederick Co. (45-1w, this paper).
8. Frederick limestone, eastern outskirts of Frederick, Md. (Stose and Stose, 1946).
9. Waynescastle Dairy, east of Greencastle, Penna. 3.5 miles on Waynesboro road (47-26w, this paper).
10. Pennsylvania Turnpike, 1 mile east of Midway Station, Bedford Co., Penna. (47-11w, this paper)
11. Tyrone, Penna., one Little Juniata River at Birmingham Station, 4 miles east of Tyrone (47-21w, this paper).
12. Ashland County, Ohio (Cohee, 1948, p. 1431).
13. Seneca County, Ohio (Cohee, 1948, p. 1431).
14. Sandusky County, Ohio (Cohee, 1948, p. 1431).
15. Geauga County, Ohio (Cohee, 1948, p. 1430).
16. Wilson Petroleum Co., No. 1 Arcade, Arcade, Wyoming Co., N.Y. (Fettke, 1948, p. 1468).
17. Lehigh Valley, Penna. (Hills, 1935, p. 124; Howell *et al.*, 1950, p. 1361–1363).
18. Northeast New Jersey (Howell, 1945).
19. Dutchess County, Pine Plains Village, N.Y. (Knopf, 1946, p. 1212).
20. Broadalbin quadrangle, N.Y. (Miller, 1911).
21. Little Falls quadrangle, N.Y. (Cushing, 1905).
22. Whitehall quadrangle, N.Y. (Wheeler, 1942, p. 523).
23. Burlington, Vt. (Cady, 1945, p. 537).
24. Highgate Gorge, Vt. (Schuchert, 1937, p. 1067).
25. Canton quadrangle, N.Y. (Chadwick, 1920, p. 217, 218).
26. Theresa quadrangle, N.Y. (Cushing *et al.*, 1910).
27. Eastern Penna., Langhorne and Durham Creek, Bucks Co. (Howell *et al.*, 1950, p. 1358–1361).

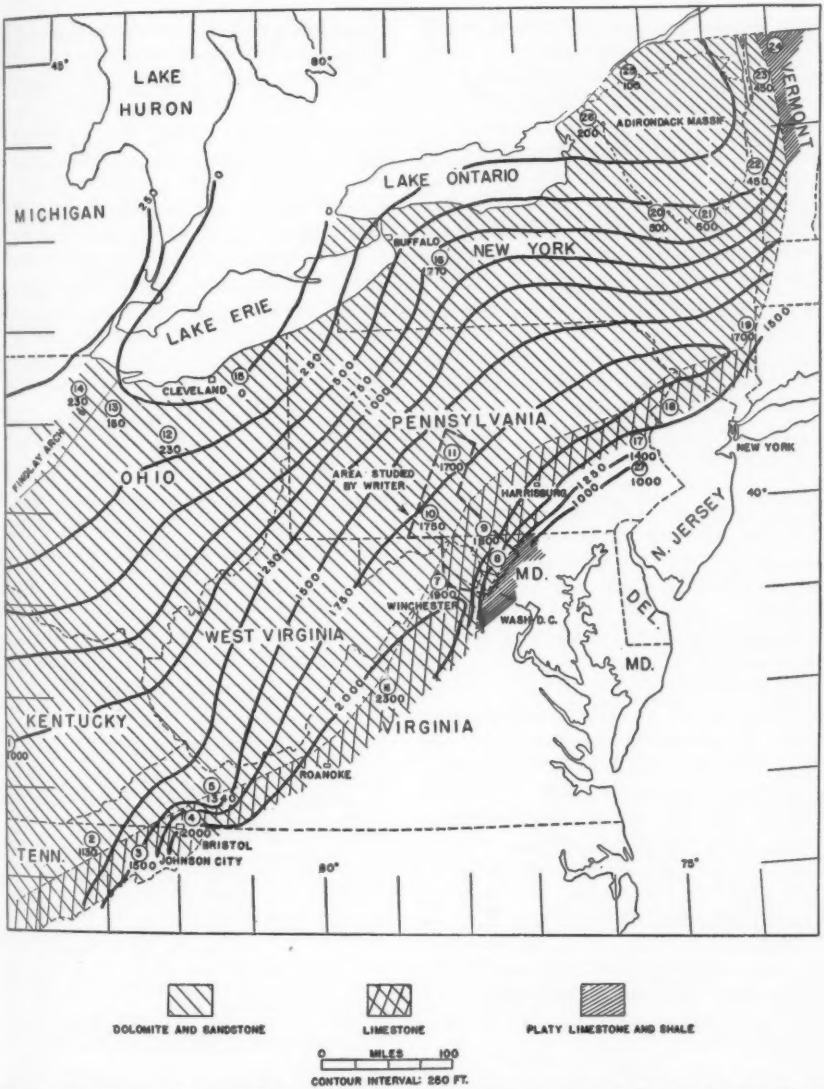


FIGURE 4.—ISOPACHOUS-FACIES MAP OF FRANCONIAN AND TREMPEALEAUIAN STAGES OF THE UPPER CAMBRIAN, NORTHEASTERN UNITED STATES

caused the biofacies change in the Franconian faunas.

The Upper Cambrian Hiatus

A faunal and lithic hiatus separates rocks of Dresbachian and Franconian age nearly everywhere over the mid-continent region (Howell and Lochman, 1939, p. 120, 121). Strata in the Appalachian geosyncline as well indicate that during latest Dresbachian and earliest Franconian time there was a period of shallower water and more clastic sedimentation.

The thick clastic sediments of the lower

sandy Gatesburg and the Big Spring Station member of the Conococheague are of middle and latest Dresbachian age (*Crepicephalus* and *Aphelaspis* faunizones). Eastward and northward in the geosyncline from the central Appalachians, even stronger evidence of this well-known hiatus exists. Fossils of Franconian age are scarcer in the Conococheague than are those of the other stages, and possibly in northeastern Pennsylvanian and New Jersey all middle Upper Cambrian beds are missing. Howell *et al.* (1950) have found no Franconian trilobites in this area. Their field evidence shows unconformable relations between the Trempealeauian Allentown formation and Dresbachian Limesport limestone—the whole Franconian seems absent (Howell *et al.*, 1950, p. 1363; Howell, 1945, p. 30). The unconformity in the Frederick Valley between the Trempealeauian Frederick limestone and Lower Cambrian sandstone (Stose and Stose, 1946) may also have been formed in part during the Dresbachian-Franconian interval.

Neither Franconian nor late Dresbachian (*Aphelaspis*) faunas have been found in Vermont although trilobites of early and middle Dresbachian age have been found in the Rockledge breccia, and many of Trempealeauian age are known from the Gorge (Raymond, 1924, 1927; Shaw, 1950). Rasetti (1944, 1945, 1946, 1948) has described numerous trilobites from the exotic Cambrian boulders of the Levi's formation in Quebec. Although Early and Medial Cambrian and early and latest Late Cambrian faunas are known, no typical Franconian trilobite genera have been recorded. Thus, whereas no actual unconformity has been seen in the Upper Cambrian of the northern Appalachian geosyncline, a considerable middle Upper Cambrian hiatus is suggested.

General Paleogeography

All the Upper Cambrian formations studied give evidence of deposition in very shallow marine water probably less than 100 feet deep over wide areas. Their cyclical sedimentation may have resulted from crustal movement, but in no respects do these deposits resemble the easterly derived tectonic sediments of the Lower Cambrian in this area. Assumedly, the border lands on both sides of the Appalachian trough were very low with little relief. Slight intermittent uplift of these lands during a period of slow marine transgression might have caused the cyclical deposition in the dolomite-orthoquartzite facies and the regular alternation of silty dolomite and limestone in the Conococheague. A small fluctuation of sea level might have shifted the shore zone scores of miles in either direction and thus radically altered the depositional environment at places many miles from shore. Yet other factors than sea level change may have caused the cycles, e.g. shifting of river mouths or regular climatic alternation.

The well-worked sediments show stable Upper Cambrian conditions with slowly transgressing seas which reached their maximum spread in latest Cambrian and earliest Ordovician times. Deposits of these ages are conformable and essentially gradational. The Gatesburg and Conococheague formations are of the miogeosynclinal type, but the Frederick limestone indicates strongly that these may have graded easterly into a eugeosynclinal belt like that of the Vermont slate sequence.

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Short Notes

PRECISION OF GEOLOGIC DATA

By Warren B. Hamilton

The concept of significant figures has long been essential in quantitative work in all fields. Geology is becoming increasingly quantitative: yet the obvious fact that data are no more accurate than the methods of their measurement is often disregarded. Many geologists present quantiative data in their true accuracy; many others do not. Numerous examples of the latter could be cited from virtually any phase of geology; the following generalizations are merely representative.

Stratigraphic thicknesses of formations studied in detail are often expressed to the nearest 5 feet or even 1 foot. This stated precision is warranted for plane-tabled sections or for well logs where structures are simple: but too often the thicknesses given are simply sums of visual field estimates or rough measurements of individual beds in formations. Each of these individual estimates may have an error of 10 or even 20 per cent, and the errors are likely to be mostly in one direction for the work of any one man. The final tabulation will be in error by the algebraic sum of the individual errors.

Where a stratigraphic thickness is pieced together from data assembled over a wide area, the total computed thickness will be put in error not only by errors in the component measurements but also by variations in thicknesses of individual units over the area.

Stratigraphic or paleontologic correlations are often attempts to establish moments of geologic time, and can thus be regarded as quantitative measurements. A much greater precision in time than is warranted is often claimed for correlations, with the result that subsequent conclusions may be misleading. Gilluly (1949) gave an excellent statement of this problem.

Chemical analyses of rocks are commonly presented with data to two decimal places, and significance is often given to slight variations. Fairbairn (Fairbairn *et al.*, 1951, p. 70) in the conclusion to a stimulating study of precision in rock analyses states that in chemical analyses of silicate rocks,

> "The retention of two figures beyond the decimal point, except for very minor constituents, becomes ridiculous for most petrologic purposes. The chemist is justified in presenting his results in this way since addition to 100 per cent is a test of his work. Petrologists, however, should ruthlessly disregard unnecessary detail beyond the decimal point."

Modal analyses of thin sections are commonly given to one decimal place, even for coarse-grained rocks. Analyses tabulated by Chayes (Fairbairn *et al.*, 1951, p. 61) show that values obtained from different thin sections from a single specimen differ enough to make delusive any reporting of more than two significant figures for major components, one for minor.

The presentation of a pseudo-precise modal analysis as representative of a mass of variable character tends to disguise the variability and to give an unwarranted inference of uniformity. Precise stratigraphic measurements have a similar effect in disguising variations in thickness.

Averages of any data are no more precise than are the assumptions under which the data are combined. Figures, still widely cited, for the "average composition of the lithosphere" (Clarke, 1924, p. 34) give values to two and three decimal places for each oxide. This average is based on the assumption (among others) that all igneous rocks, continental and oceanic, have the average composition shown by the statistical average of rock analyses: yet only a small fraction of the analyses are from the quantitatively most important oceanic areas.

Round estimates are often changed into figures of precise appearance when conversion is made into other units. Thus 1000 feet expressed

as 305 meters suggests an accuracy not present in the initial estimate.

The increasing use of quantitative data in geology is heartening and necessary: but it is important that it be remembered that the data are of no greater accuracy than is permitted by the methods upon which they are based. Methods of measurement should be analyzed to demonstrate their accuracy, and the results of measurements should be presented in their true degree of precision.

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